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RELATIONS BETWEEN METAMORPHISM AND MAGMATISM NEAR CHETICAMP, CAPE BRETON ISLAND, NOVA SCOTIA

K.L. Currie

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RELATIONS BETWEEN METAMORPHISM AND MAGMATISM NEAR CHÉTICAMP, CAPE BRETON ISLAND, NOVA SCOTIA

Abstract

The oldest rocks of the Chéticamp region form nebulous schlieren and zones of retrograded granulites within granitoid gneisses (Pleasant Bay complex). The complex contains a tectonically bounded lenticle of anorthosite. The Pleasant Bay complex had been migmatized and intruded by basic dykes prior to emplacement of the granodioritic Cheticamp pluton about 550 Ma at the culmination of deformation and metamorphism with peak conditions near 650°C, 900 MPa (9.0 kilobars) At about 440 Ma older rocks were unconformably overlain by a bimodal basalt-rhyolite suite, and younger sedimentary rocks of the Jumping Brook complex. This complex was metamorphosed and deformed during Silurian and Devonian time reaching temperature-pressure conditions up to 650°C, 350 MPa (3.5 kilobars). Magmatic activity accompanied metamorphism, terminated by emplacement of high level granite plutons and correlative bimodal basalt-rhyolite volcanics of the Late Devonian Fisset Brook Formation. Immediately following magmatic activity, talus, fanglomerate and arkosic sand of the Horton Group collected at the foot of scarps. Carbonates and evaporites of the Mississippian Windsor Group accumulated in closed basins. Renewed clastic sedimentation in Pennsylvanian time produced sandstones of the Canso and Riversdale groups.

The Chéticamp region forms part of an unusual Avalonian terrane with a Grenvillian basement. The association of Grenvillian basement with Avalonian magmatism can be reconciled if the Late Precambrian-Early Paleozoic history of this region involved repeated breakup and re-assembly of the continental margin. The association of mylonites, thrusts and transcurrent faults with an intricate sequence of sedimentary facies suggests major transcurrent motion during Carboniferous time.

Résumé

Les roches les plus anciennes de la région de Cheticamp forment des schlierens et des zones de granulites régressifs au sein de gneiss granitoïdes (complexe de Pleasant Bay). Le complexe contient une lentille d'anorthosite dont les limites sont d'origine tectonique. Le complexe de Pleasant Bay avait été migmatisé et coupé par des dykes basiques avant la mise en place du pluton granodioritique de Cheticamp, il y a environ 550 Ma, à l'apogée de la période de déformation et de métamorphisme, lorsque les conditions ont atteint près de 650°C et 900 MPa (9,0 kilobars). Les roches plus anciennes ont été recouvertes en discordance par une suite basaltique-rhyolitique bimodale et par les roches sédimentaires plus récentes du complexe de Jumping Brook, il y a environ 440 Ma. Ce complexe a été métamorphisé et déformé au cours du Silurien et du Dévonien, la température et la pression ayant atteint 650°C et 350 MPa (3,5 kilobars). Le métamorphisme a été accompagné par une activité magmatique qui a abouti à la mise en place de plutons à des niveaux élevés dans la colonne lithostratigraphique et des roches volcaniques basaltiques-rhyolitiques corrélatives de la formation de Fisset Brook du Dévonien supérieur. Cette activité magmatique a été immédiatement suivie par l'accumulation, au pied des talus, des cônes d'éboulis, des fanglomérats et du sable arkosique du groupe de Horton. Les roches carbonatées et les évaporites du groupe mississippien de Windsor se sont accumulées dans des bassins fermés. Une nouvelle sédimentation clastique au cours du Pennsylvanien a produit les grès des groupes de Canso et de Riverdale.

La région de Cheticamp fait partie d'un terrain avalonien insolite au socle grenvillien. On peut concilier l'association d'un socle grenvillien et d'un magmatisme avalonien si l'on suppose que la marge continentale de la région a été fracturée et rassemblée à maintes reprises au cours de la période allant de la fin du Précambrien au début de Paléozoïque. L'association de mylonites, de failles chevauchantes et de décrochements et d'une séquence complexe de faciès sédimentaires laisse croire qu'un important mouvement transversal s'est produit au cours du Carbonifère.

SUMMARY

Rocks of the Chéticamp region fall into six major divisions, namely (1) polymetamorphic, polydeformed gneisses of the Pleasant Bay complex, (2) the granodioritic Cheticamp pluton of Late Precambrian-Cambrian age, (3) a metamorphosed volcano-sedimentary assemblage of Silurian(?) age, (4) a varied magmatic assemblage of Devono-Carboniferous age, including bimodal volcanics and high level plutons, (5) a varied Carboniferous sedimentary section, and (6) Carboniferous and older cataclastic rocks of uncertain protolith.

The oldest parts of the Pleasant Bay complex form zones and schlieren of retrogressed granulites within a younger granitoid gneiss. Granulite facies metamorphism also occurs associated with a tectonic slice of massif anorthosite. These antique parts of the complex appear to be remnants of Grenvillian basement. Other parts of the Pleasant Bay complex include quartzofeldspathic gneisses

SOMMAIRE

Les roches de la région de Cheticamp se groupent en six grandes divisions, soit (1) les gneiss polymétamorphiques et polydéformés du complexe de Pleasant Bay, (2) le pluton granodioritique de Cheticamp, du Précambrien-Cambrien inférieur, (3) une association volcano-sédimentaire métamorphisée d'âge silurien(?), (4) une association magmatique variée dévono-carbonifère qui comprend des roches volcaniques bimodales et des plutons cristallisés à un niveau élevé dans la colonne lithostratigraphique (5) une section sédimentaire carbonifère variée et (6) des roches carbonifères et des roches clastiques plus anciennes provenant de roches mères incertaines.

Les parties les plus anciennes du complexe de Pleasant Bay forment des zones et des schlierens de granulites régressifs dans un gneiss granitoïde plus récent. Un métamorphisme ayant atteint le faciès des granulites est également associé à une tranche tectonique d'un massif d'anorthosite. Ces parties anciennes du complexe seraient des restes du socle grenvillien. D'autres parties du complexe de Pleasant Bay contiennent des deduced to represent intermediate volcanic rocks metamorphosed at conditions near 650°C and 900 MPa (9.0 kilobars). These rocks may represent remnants of an Hadrynian cover sequence on basement.

The Pleasant Bay complex had been migmatized and intruded by three generations of mafic dykes prior to emplacement of the Cheticamp pluton about 550 Ma. This elongate pluton varies from diorite to two-mica leucogranite, but tends to be granodiorite. The great variation in composition may result from incorporation of peraluminous host rocks.

Older rocks were overlain by a volcano-sedimentary suite dated by zircons from rhyolitic dykes at 439 Ma. The basal volcanic suite is bimodal, and may be correlative to some of the younger gneissic granites of the Pleasant Bay complex. The volcanic rocks are overlain by pelitic and psammitic rocks, now converted to spectacular knotted schists. The volcano-sedimentary sequence shows a fairly complete sequence of progressive metamorphism from subgreenschist facies (diaspore-bearing) up to peak conditions of 650°C, 350 MPa (3.5 kilobars). A variety of synmetamorphic granitoid rocks include gneissic leucogranite, lineated muscovite pegmatites, and crosscutting to concordant massive pegmatites. Biotite dates suggest metamorphism was of Early Devonian age.

Bimodal, post-tectonic magmatic activity emplaced large, high level potassic granites, and basalt-rhyolite volcanics of the Late Devonian to Early Mississippian Fisset Brook Formation.

The Pleasant Bay complex exhibits generally north to north-northwest trending, steeply plunging isoclinal folds, only moderately modified by later folding. The Jumping Brook complex was folded into overturned and recumbent north-northeast trending folds prior to significant metamorphism. During metamorphism the rocks were refolded into tight, north-trending and plunging, upright folds. All rocks older than Pennsylvanian are affected by intense mylonitization concentrated in north-northeast trending zones along the sea shore. Mylonites appear to have formed at several times, most recently in Mississippian time when thrusting to the west stacked metamorphic zones, leading to the present pattern in which progressively higher grade rocks are exposed from west to east. The association of thrust and transcurrent faulting during Carboniferous time together with an intricate sequence of sedimentary facies, suggests major transcurrent motion at this time. Mylonite zones of this age now dominate the appearance of the rocks.

Volcanic rocks of the Jumping Brook complex display a low tenor of Cu mineralization in the mafic volcanics and Pb-Zn-Ag mineralization in the acid volcanics. Remobilization, presumably by circulating solutions, has created small enriched zones along faults. Many of these faults appear to be mineralized near the unconformity below the Carboniferous sedimentary rocks. Mobilization by folding and faulting locally thickened gypsum horizons of the Windsor Group into mineable deposits, and created numerous gypsum-calcite veins with a low tenor of Cu or Pb-Zn.

gneiss quartzofeldspathiques qui représenteraient des roches volcaniques intermédiaires métamorphisées à une température de près de 650°C et une pression de près de 900 MPa (9,0 kilobars). Ces roches pourraient représenter des restes d'une séquence de couverture hadrinienne sus-jacente au socle.

Le complexe de Pleasant Bay avait été migmatisé et coupé par trois générations de dykes mafiques avant la mise en place du pluton de Cheticamp, il y a environ 550 Ma. Ce pluton allogé passe d'une diorite à un leucogranite à deux micas, mais en général, sa composition est celle d'un granodiorite. La variation marquée de la composition pourrait être le résultat de l'incorporation de roches mères hyperalumineuses.

Des roches plus anciennes ont été recouvertes par une suite volcano-sédimentaire qui date de 439 Ma; l'âge est fondé su des zircons extraits de dykes rhyolitiques. La suite volcanique basale est de nature bimodale et pourrait être mise en corrélation avec certains granito-gneiss plus récents du complexe de Pleasant Bay. Les roches volcaniques sont sous-jacentes à des roches pélitiques et psammitiques maintenant transformées en de spectaculaires schistes tachetés. La suite volcano-sédimentaire présente une séquence plus ou moins complète produite par métamorphisme progressif et allant du faciès inférieur à celui des schistes verts (qui contient du diaspore) à des conditions culminant à 650°C et 350 MPa (3,5 kilobars). Du leucogranite gneissique, des pegmatites linéaires à muscovite et des pegmatites massifs entrecroisés ou concordants font partie des roches granitoides synmétamorphiques. La datation de la biotite laisse croire que le métamorphisme a eu lieu au début du Dévonien.

Une activité magmatique bimodale, post-tectonique, a mis en place de vastes granites potassiques cristallisés à un niveau élevé dans la colonne lithostratigraphique, ainsi que les roches volcaniques basaltiques-rhyolitiques de la formation de Fisset Brook du Dévonien supérieur au Mississippien inférieur.

Le complexe de Pleasant Bay présente des plis isoclinaux, généralement à plongement brusque et à orientation nord ou nord-nord-ouest, qui n'ont été que très peu modifiés aux cours d'étapes subséquentes de plissement. La formation de plis retournés et couchés à orientation nord- nord-ouest dans le complexe de Jumping Brook a précédé le métamorphisme intense au cours duquel les roches ont été déformées à nouveau pour former des plis droits serrés, à orientation et à plongement nord. Toutes les roches pré-pennsylvaniennes ont été touchées par une mylonitisation intense concentrée dans des zones à orientation nord-nord-est le long du littoral. Les mylonites semblent s'être formées à différents moments, les plus récentes datant du Mississippien, lorsqu'un chevauchement vers l'ouest a empilé les zones métamorphiques, produisant le motif actuel dans lequel des roches de métamorphisme progressivement plus intense affleurent de l'ouest vers l'est. La formation de failles chevauchantes et de décrochements au cours du Carbonifère et la présence d'une séquence de faciès sédimentaires laissent croire qu'un important mouvement transversal s'est produit au cours de cette période. De nos jours, les roches sont dominées par des zones de mylonite carbonifère.

Les roches volcaniques mafiques du complexe de Jumping Brook ont une faible teneur en cuivre, tandis que les roches volcaniques acides contiennent du plomb, du zinc et de l'argent. Une nouvelle mobilisation chimique, vraisemblablement provoquée par des solutions en circulation, a donné lieu à des petites zones enrichies le long des failles. Un grand nombre de ces failles semblent avoir été minéralisées près de la discordance située sous les roches sédimentaires carbonifères. La mobilisation provoquée par la formation de plis et de failles a épaissi certains horizons de gypse du groupe de Windsor pour former des gisements exploitables et a créé de nombreux filons de gypse et de calcaire qui contiennent de faibles teneurs en cuivre ou en plomb-zinc.

Analysis of bulk rock geochemistry demonstrates marked differences between metamporphosed supracrustal rocks of the Pleasant Bay and Jumping Brook complexes. The former appear to be derived from volcanic, or volcanogenic material, as indicated by a discriminant function of Shaw (1972), with intercalations of limy material. Under the conditions of metamorphism reached by the Pleasant Bay complex, the limy bands produce the assemblage diopside-scapolite-grossular ± hornblende ± epidote. Thermodynamic and experimental data show that this assemblage is stable only at high temperatures and pressures (greater than 650°C and 800 MPa or 8 kilobars). Since this assemblage appears intermittently throughout the area, the Pleasant Bay complex probably was metamorphosed throughout to these conditions. Indirect evidence suggests that this metamorphism took place in Late Precambrian time. By contrast, the Jumping Brook complex contains salic and mafic volcanics (basalt and rhyolite) but lacks intermediate volcanic rocks. Limy bands are absent, but pelitic compositions, which are very rare or absent in the Pleasant Bay complex, are abundant. The characteristic sequence of porphyroblasts diaspore-chloritoid-garnet-biotite-staurolite-cordierite-andalusite(-sillimanite) shows that a progressive metamorphic sequence is preserved. Application of published geothermobarometers established a well-defined P-T trajectory which passes through a corner of the stability field for kyanite. Rare relicts of kyanite are found in staurolite-grade schists.

Both periods of metamorphism generated granitoid rocks in the form of nebulitic gneisses, lineated pegmatites, and pegmatitic pods of presumed anatectic origin. However the major granitoid plutonism associated with both periods of metamorphism clearly postdated the peak of metamorphism. In the case of the Pleasant Bay complex, this later material (Cheticamp pluton) is clearly of deep-seated origin (I-type pluton), although it probably has been contaminated by hybridization with or anatexis of surrounding host rocks. In the case of the Jumping Brook complex, the post-tectonic granites (Margaree and Salmon Pool plutons) are high level, potassic, mildly peraluminous granites without associated mafic phases, whose protolith is uncertain, but may be crustal.

Rocks of the Chéticamp region correlate well with other parts of the Cape Breton Highlands, and less certainly with the Avalon and Gander zones of Newfoundland. The presence of the 550 Ma Cheticamp pluton strongly suggests that this region forms part of an Avalonian terrane, but the presence of extensive Siluro-Ordovician and Devonian magmatism suggests the Gander zone, while the occurrence of massif-type anorthosite suggests a Grenvillian basement. These apparently contradictory indications can be reconciled if the Late Precambrian and Early Paleozoic history of this region involved repeated breakup and reassembly of the continental margin. The nature of the Pleasant Bay complex implies metamorphism associated with westward directed subduction during Late Precambrian time. The nature of the Siluro-Devonian event remains unclear. Possibly it represents rifting after Taconic orogeny. The present mineralogy and texture of the rocks result principally from Devonian metamorphism, which locally reached upper amphibolite grade, and associated deformation. Petrographic and geophysical data suggest post-Devonian history of the Cape Breton Highlands has been dominated by large scale transcurrent motions.

L'analyses de la géochimie de la roche entière révèle l'existence de différences marquées entre les roches supracorticales métamorphisées du complexe de Pleasant Bay et celles du complexe de Jumping Brook. Les premières seraient dérivées de matériau volcanique ou d'origine volcanique, comme l'indique la fonction discriminante de Shaw (1972), avec des intercalations de matériau calcaire. Dans les conditions de métamorphisme atteintes par le complexe de Pleasant Bay, les bandes calcaires produisent l'association diopside-scapolite-grossulaire+hornblende+ épidote. Les données thermodynamiques et expérimentales montrent que cette association n'est stable qu'à des températures et des pressions élevées (supérieures à 650°C et à 800 MPa ou 8 kilobars). Puisque cette association se retrouve de façon intermittente partout dans la région, le complexe de Pleasant Bay aurait donc été métamorphisé partout dans de pareilles conditions. Des indices indirects portent à croire que ce métamorphisme a eu lieu à la fin du Précambrien. Par opposition, le complexe de Jumping Brook contient des roches volcaniques saliques et mafiques (basalte et rhyolite), mais est dépourvu de roches volcaniques intermédiaires. Les bandes calcaires y sont absentes tandis que les zones pélitiques, très rares ou absentes dans le complexe de Pleasant Bay, y abondent. La présence d'une séquence caractéristique de porphyroblastes, soit diaspore-chloritoide-grenat-biotite-staurotide-cordiérite-andalousite(-sillimanite), indique qu'il y a une séquence métamorphique progressive préservée. L'utilisation de géothermobaromètres a permis d'établir un tracé P-T bien défini, qui traverse un coin du domaine de stabilité de la cyanite. De rares résidus de cyanite se trouvent dans les schistes à staurotide.

Les deux périodes de métamorphisme ont donné lieu à des roches granitoides sous forme de gneiss nébulitiques, de pegmatites linéaires et de masses pegmatitiques allongées d'origine vraisemblablement anatectique. Cependant, il est évident que la plus grande partie du plutonisme granitoide associé aux deux périodes de métamorphisme a eu lieu après la culmination du métamorphisme. Dans le cas du complexe de Pleasant Bay, le matériau plus récent (pluton de Cheticamp) est manifestement d'origine profonde (pluton de type I), bien qu'il ait probablement été contaminé par hybridation avec les roches mères encaissantes ou par anatexie de ces roches mères. Dans le cas du complexe de Jumping Brook, les granites post-tectoniques (plutons de Margare et de Salmon Pool) sont des granites potassiques, légèrement hyperalumineux, cristallisés à des niveaux élevés dans la colonne lithostratigraphique et dépourvus de phases mafiques; l'origine des roches mères est incertaine mais pourrait être crustale.

Une bonne corrélation a été établie entre les roches de la région de Cheticamp et celles d'autres parties des hautes-terres du Cap-Breton; la corrélation est moins certaine avec les roches des zones d'Avalon et de Gander, à Terre-Neuve. La présence du pluton de Cheticamp, qui date de 550 Ma, porte fortement à croire que cette région fait partie d'un terrain avalonien; toutefois la présence, sur de vastes étendues, d'un magmatisme siluro-ordovicien et dévonien, laisse croire qu'elle fait partie de la zone de Gander, tandis que la présence d'un massif d'anorthosite porte à croire qu'il s'agit d'un socle grenvillien. On peut concilier ces indications apparemment contradictoires si l'on suppose que la marge continentale de la région a été fracturée et rassemblée à maintes reprises au cours de la période allant de la fin du Précambrien au début du Paléozoique. La nature du complexe de Pleasant Bay laisse supposer qu'il y a eu métamorphisme associé à une subduction en direction ouest au cours du Précambrien tardif. La nature de l'évènement siluro-dévonien n'est par évidente; il pourrait s'agir de la formation de fissures suivant l'orogénèse du Taconique. La minéralogie et la texture actuelles des roches sont dues principalement au métamorphisme dévonien qui a, par endroits, atteint le faciès supérieur des amphibolites et la déformation connexe. Les données pétrographiques et géophysiques laissent croire que l'histoire post-dévonienne des hautes-terres du Cap-Breton a été dominée par des mouvements transversaux de grande envergure.

INTRODUCTION

General statement

The Cape Breton Highlands form an elevated block of metamorphic and plutonic igneous rocks with small outliers of Carboniferous sedimentary and volcanic rocks around its margins. Neither the internal stratigraphy nor the correlation with other parts of the Canadian Appalachians has been established. Reconnaissance mapping depicted the bulk of the highlands as Precambrian in age (Fletcher, 1885; MacLaren, 1955; Neale, 1963), although MacLaren (1955) and Neale (1963) recognized the presence of Devonian igneous rocks. Wiebe (1972) worked out a detailed succession of events along the east side of the Cape Breton Highlands which involved several periods of deposition, deformation and plutonism producing complex deformed and metamorphosed basement-cover relations. Neale and Kennedy (1975) and Keppie (1979) expressed views similar to Wiebe's, namely that two or more ages of supracrustal rocks occur, at least locally, in the Cape Breton Highlands, and that basement-cover relations could be worked out. By contrast, MacDonald and Smith (1980) held that only a single protolith occurred in the northeastern corner of the highlands, although it has been variably metamorphosed. They further held that only two significant ages of granitoid generation and emplacement could be recognized, at least in the area studied by them. Similar views have been advanced for the southern highlands by Jamieson and Craw (1983).

Various authors expressed equally diverse views on the correlation of the rocks of the Cape Breton Highlands with other parts of the Canadian Appalachians. Neale and Kennedy (1975) and MacDonald and Smith (1980) correlated highland rocks with southwestern Newfoundland, whereas Kennedy (1975) suggested correlation with the Gander zone of Newfoundland, and Keppie (1979) thought they might correlate with offshore rocks to the east of Newfoundland. A more traditional, though hardly more convincing, correlation linked the Cape Breton Highlands to southeastern Cape Breton Island (Milligan, 1970; Helmstaedt and Tella, 1973).

These divergent and mutually contradictory views arise largely from lack of information. Large parts of the Cape Breton Highlands have not been mapped in detail, particularly the western part of the highlands. Prior to this study, only Wiebe (1972) had attempted to unravel the complexities of the gneisses, and he had abandoned the attempt on the more complex migmatitic rocks (Coastal gneisses). The present study concentrated on a little known region of the Cape Breton Highlands and attempted to extract all possible information from the relations of granitoid and metamorphic rocks, particularly small scale relations in metamorphic complexes. The study aimed to identify igneous and metamorphic components of such complexes, and determine the relations between them. The larger scale aim was to integrate this information with previous studies to produce a more detailed and unified picture of the development of the Cape Breton Highlands, and improved correlations with other regions.

Location and access

The mapped area consists of a roughly triangular region along the northwestern shore of Cape Breton Island, depicted on map sheet 11K of the National Topographic System and bounded by 46°32' and 46°50' north latitude, and 60°40" and 61°05' west longitude (Fig. 1). About 80% of this area lies within Cape Breton Highlands National Park. The towns of Chéticamp and Pleasant Bay lie near the southwest and northwest corners of the area respectively. The Cabot Trail, a paved scenic highway which winds along the western and northern sides of the area, forms the principal access route. The sea cliffs, particularly north of Jumping Brook, are accessible by boat only, whereas the interior of the area can be reached with varying degrees of difficulty by foot trails within the park and by bush roads and trails outside the park. Most of these roads and trails follow the major streams and commonly end abruptly in difficult terrane. The central parts of the area can be reached only by helicopter or backpacking.

The amount of outcrop decreases rapidly away from the sea coast, except in the valleys of major streams. The high barrens near the eastern edge of the area are virtually devoid of outcrop. I examined in detail the sea shore, roadcuts along the Cabot Trail, the valleys of major streams, and foot paths and bush roads, with widely spaced traverses across the intervening ground. Since the streams tend to cut almost perpendicularly across the strike of the rocks, this plan gave excellent cross-sections. Fieldwork occupied six weeks in 1974, three weeks in 1975 and six weeks in 1976. Laboratory work commenced in 1975 and was completed in 1980.

Physiography and glacial geology

The mapped area consists of a deeply dissected, rolling plateau. The relief, both local and total, approaches 500 m. Along the coast the plateau drops off abruptly in a series of spectacular cliffs, either directly into the sea, or to a narrow coastal plain underlain by Carboniferous strata. Major streams have cut into the cliffs for distances varying with the size of the stream, so that the relatively small Jerome Brook has cut back only a few hundred metres, Corney Brook has incised about 2 km, and the largest streams, Chéticamp, Mackenzie, and North Aspy rivers, have achieved relatively mature profiles. All major streams follow linear courses, commonly with sharp elbows, and appear to be structurally controlled. In the lower, incised, parts of the streams outcrop is rare. The lower course of the major streams contains much fluvial sand and gravel. Mackenzie River, Corney Brook and Chéticamp River are all dissecting old river mouth deposits as much as 25 m above sea level, implying either very recent uplift or changes in the hydraulic regime of the streams.

The lower parts of the area support a moderate growth of scrubby spruce and balsam forest, devastated in most areas by spruce budworm. The parts of the region above about 400 m are essentially barren, underlain by gently rolling muskeg with many string bogs, small ponds and meandering streams.

Evidence of glaciation is not prominent. Glacial polish and striation occur on a few outcrops along the Cabot Trail where the top of the outcrop has been exposed and cleaned by road maintenance operations. No reliable indication of ice movement direction was found, and constructive glacial forms appear to be virtually absent. Saprolite, presumably relict from deep pre-glacial weathering, and implying little or no glacial abrasion, occurs a few kilometres south of the mapped area (McKeague et al., 1983).

Acknowledgments

This study was undertaken with the co-operation of Parks Canada and Cape Breton Highlands National Park. At various times the project was discussed with Robert Wiebe of Franklin and Marshall College, Duncan Keppie and Peter Giles of the Nova Scotia Department of Mines and Energy, and Rebecca Jamieson of Dalhousie University. N.A.C. Rey, Geological Survey of Canada, did much of the preliminary petrographic work and wrote drafts on the relations of metamorphism and deformation. G.J. Pringle, Geological Survey of Canada, made the electron probe analyses, as well as many supplementary petrographic observations. The debt to the previous mapping by E.R.W. Neale and the late A.S. MacLaren of the Geological Survey of Canada will be obvious to all; the present study could not have been conceived without it. The manuscript has benefited greatly from thorough critical reading by W.H. Poole, Geological Survey of Canada, and R.A. Jamieson, Dalhousie University.

DESCRIPTION OF MAP UNITS

Introduction

Most rocks in the Chéticamp region fall naturally into one of six major divisions, namely: (1) biotitic, amphibolitic and granitoid gneisses of the Pleasant Bay complex; (2) the granodioritic Cheticamp pluton; (3) volcanic, pelitic and psammitic schists of the Jumping Brook complex; (4) leucocratic granitic rocks, including the Margaree pluton and Salmon Pool granite; (5) a varied and important Carboniferous section including the Horton, Windsor, Canso, and Riversdale groups; and (6) distinctive cataclastic rocks whose protolith in many cases cannot be certainly identified. In addition to these major divisions, mafic rocks occur whose position is uncertain. Contacts among the plutonic rocks tend to be poorly exposed, complex, ambiguous and strongly deformed. The Pleasant Bay complex appears to contain an old protolith which was intruded by the Cheticamp pluton. The Jumping Brook complex probably unconformably overlies the Cheticamp pluton, but has been metamorphically welded to the Pleasant Bay complex before or during emplacement

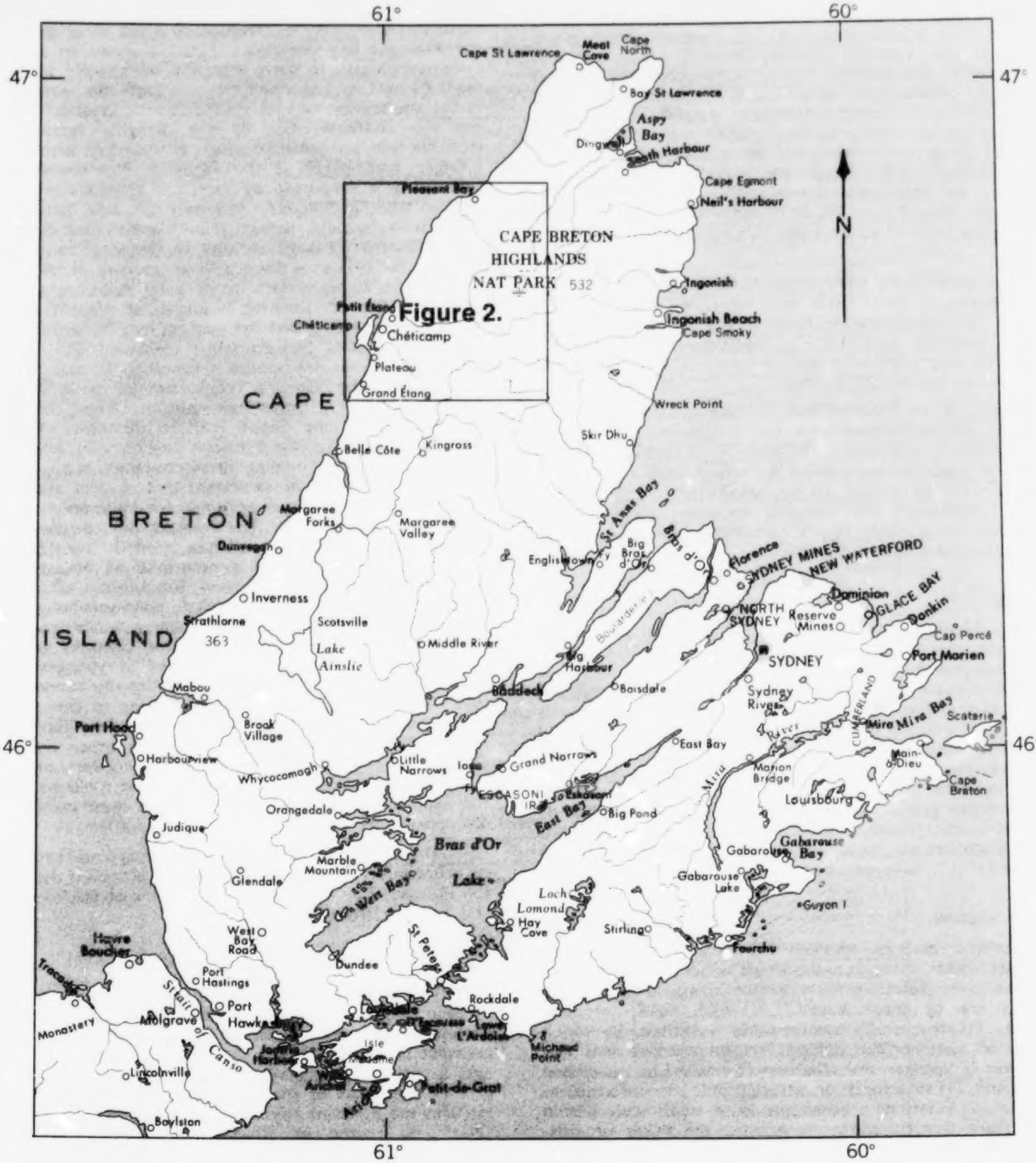
61°
60°
47°
46°
Cape St Lawrence
Meat Cove
Cape North
Bay St Lawrence
Aspy Bay
Dingwall
South Harbour
Cape Egmont
Neil's Harbour
N
Pleasant Bay
CAPE BRETON HIGHLANDS NAT PARK
532
Ingonish
Ingonish Beach
Cape Smoky
Petit Étang
Figure 2.
Chéticamp I
Chéticamp
Plateau
Grand Étang
Wreck Point
CAPE
BRETON
ISLAND
Belle Côte
Kingross
Skir Dhu
St Anns Bay
Margaree Forks
Margaree Valley
Dunvegan
Big Bras d'Or
Bras d'Or
Englishtown
Florence
SYDNEY MINES
NEW WATERFORD
Inverness
Dominion
GLACE BAY
Donkin
NORTH SYDNEY
Reserve Mines
Strathlorne
363
Scotsville
Lake Ainslie
Middle River
Boularderie
Cap Percé
Port Morien
SYDNEY
Big Harbour
Baddeck
Boisdale
Sydney River
Mabou
Port Hood
Brook Village
Mira
Mira Bay
Scatarie
CUMBERLAND
East Bay
Harbourview
Whycocomagh
Little Narrows
Iona
Grand Narrows
Main-à-Dieu
Cape Breton
Marion Bridge
ESKASONI IR
Eskasoni
East Bay
Big Pond
Louisbourg
Orangedale
Judique
Bras d'Or
Lake
Gabarouse
Gabarouse Bay
Marble Mountain
West Bay
Gabarouse Lake
Glendale
Loch Lomond
Hay Cove
Guyon I
Stirling
Havre Boucher
West Bay Road
Fourchu
Tracadie
Dundee
St Peters
Port Hastings
Strait of Canso
Rockdale
Port Hawkesbury
Louisdale
St Peters
Lower L'Ardoise
Monastery
Mulgrave
Isle Madame
Michaud Point
Janvrin Harbour
West Arichat
Lincolnville
Arichat
Petit-de-Grat
Boylston

Figure 1. Location map of study area. (GSC 188584)

of the Margaree pluton and Salmon Pool granite. The Carboniferous section rests with great unconformity on older rocks, but strata as young as the Mississippian Windsor Group have been strongly deformed by folding and cataclasis. These relationships, based mainly on observations of small-scale cross-cutting relationships, are summarized in the Table of Formations (Table 1).

Pleasant Bay complex

Definition and distribution

The Pleasant Bay complex includes all rocks which show prominent gneissic banding, that is small scale metamorphic layering. The complex has four end members which grade into one another, namely: strongly foliated, weakly gneissic, biotite augen granodiorite (**Hpbg**); schistose, more or less granitized, biotite-rich gneiss, locally calcareous (**Hpbp**); strongly foliated and/or lineated amphibolitic gneiss, commonly potash metasomatized (**Hpba**); and heterogeneous lit-par-lit gneiss which tends to form a matrix for the other units (unit **Hpb**). The Pleasant Bay complex locally contains significant but unmappable amounts of granitic units similar to units Hc, **SDg**, **Dl** and **Dsp** intimately associated with the gneiss in the form of schliers and thin sills, or more rarely as dykes.

In a preliminary publication (Currie, 1975) I included a small amount of anorthosite (unit Hrra) with the gneissic units to form the Grande Anse complex. Further study suggests that the anorthosite is mechanically, rather than genetically associated with the gneisses. The name Grande Anse complex has been abandoned.

Rocks of the Pleasant Bay complex outcrop in a north-south belt across the map area, and in two smaller, roughly parallel belts to the west. The complex is particularly well exposed in roadcuts along the Cabot Trail to the south and east of the village of Pleasant Bay after which the complex is named. Although the complex exhibits a variety of lithologies, the gneissic, highly metamorphosed character of the rocks imposes an overall unity.

Lithology

The Pleasant Bay complex forms a typical medium to high grade gneiss complex, characterized by small scale lithological and structural diversity. For mapping purposes a four-fold division can be applied by using the arbitrary criteria that gneissic rocks belong to the biotite gneiss unit (**Hpbp**) if the outcrop contains more than 35 per cent biotite gneiss and schist or recognizably metasedimentary rocks (quartzite, marble), to the amphibolite unit (**Hpba**) if the outcrop contains more than 50 per cent amphibolite, to the gneissic augen granodiorite unit (**Hpbg**) if the outcrop is of granitoid character and contains less than 25 per cent of non-granitoid schliers and inclusions, and to the composite gneiss unit (**Hpb**) if it falls in none of the these categories.

Contact relations

Contact relations between the Pleasant Bay complex and other units comprise the most direct evidence for basement-cover relations in the Chéticamp area. These relations are of three kinds: (1) small scale intrusive contacts, (2) structurally conformable relations in which textural or metamorphic mineral evidence shows that one component is younger than the rest (for example transposed dykes), and (3) structural or stratigraphic unconformities. Cross-cutting relations are common on a small scale within the Pleasant Bay complex. In general the dykes or sills, which range up to 30 cm across, can be readily correlated with younger granitoid units on the basis of mineralogy and texture. Figure 3 shows a typical example, in this case a dyke of the Cheticamp pluton cutting granodiorite gneiss (**Hpbg**) of the Pleasant Bay complex. Both dyke and host have been slightly remobilized resulting in boudinage in the centre right part of the photograph. A more complex example (Fig. 4) shows a folded dyke of the Cheticamp pluton cutting amphibolitic gneiss (**Hpba**). Not only has the dyke been folded, but partial remelting has probably occurred, as suggested by the prominent quartzofeldspathic rims on the dyke. Analogous relations demonstrate that granite gneiss of units Hc, **SDg**, **Dl**, **Dm** and **Dsp** invade the Pleasant Bay complex, but only units Hc and **SDg** have been deformed and metamorphosed with it. Transposed dykes occur commonly in the Pleasant Bay complex. Figure 5 shows an amphibolite dyke now parallel to the gneissosity, but readily distinguished from its host by lesser deformation (note the many lenticles in the gneiss) and lack of compositional banding. This dyke has the characteristics of the Jumping Brook complex (foliated but not gneissic) which is therefore deduced to be younger than parts of the Pleasant Bay complex. This conclusion is supported by rare but spectacular exposures such as those in Figure 6. Intensely schistose Jumping Brook complex to the right rests on granite gneiss (unit **Hpbg**) which has developed cleavage parallel to the schistosity near the contact, but retains a discordant gneissosity in the left-hand part of the figure. Such large scale relations tend to be ambiguous because shearing is invariably present along such contacts, suggesting that the contact may be tectonic, rather than stratigraphic. However when combined with small-scale evidence such as transposed dykes (Fig. 5), such exposures suggest that the Jumping Brook complex must be younger than parts of the Pleasant Bay complex. Along Fishing Cove River and along the Cabot Trail to the east, enclaves of pelitic schist within the Pleasant Bay complex appear to be correlative to the Jumping Brook complex, suggesting that final remobilization or structural assembly of the Pleasant Bay complex postdated or accompanied metamorphism of the Jumping Brook complex. The Pleasant Bay complex contains two older suites of transposed dykes, readily identified by the presence of one or two generations of potash feldspar porphyroblasts (augen) in hand specimen. The younger generation contains pink euhedral porphyroblasts up to a centimetre across in a fairly homogeneous amphibolitic matrix, whereas the older generation contains granulated, rounded augen porphyroblasts as well as younger euhedral porphyroblasts in a gneissic, compositionally banded, matrix (see Fig. 26). I interpret these relations to demonstrate a polymetamorphic character for the Pleasant Bay complex such that a component older than any other unit in the Chéticamp region is present, but a number of younger components also occur, probably including rocks as young as the Salmon Pool granite. The nature of these components is discussed under Metamorphism and Magmatism.

The Upper Devonian and/or Mississippian Horton Group rests unconformably upon, and contains abundant debris from, the Pleasant Bay complex at the mouths of Mackenzie River and Jumping Brook.

Age

The foregoing discussion shows that an "age" for the Pleasant Bay complex must be defined with care, since components of several ages are undoubtedly present. Perhaps greatest interest attaches to the oldest component, whose age is presently unknown. This age, however, must be greater than that of the Cheticamp pluton (unit **Hc**) which intrudes the Pleasant Bay complex. Cormier (1972) dated the Cheticamp pluton by Rb-Sr isochron at 530 ± 44 Ma[1],

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[1] All dates recalculated to the decay constants recommended by Steiger and Jäger (1977).

Table 1. Table of Formations

Eon	Era	Period	Group or complex	Unit*	Lithology
	Cenozoic			Qt	drift, till, fluvial deposits
	- unconformity -				
Phanerozoic	Paleozoic	Pennsylvanian	Riversdale Group	PRs	grey, flaggy sandstone
		- conformable contact -			
		Pennsylvanian/ Mississippian	Canso Group	PMCs	buff sandstone, minor red shale
		- unconformity(?) -			
		Mississippian and older		Mbr	breccia, mylonite, flaser gneiss
		- tectonic contact -			
		Mississippian	Windsor Group	MWg	limestone, dolostone, gypsum, minor red shale
		- tectonic contact -			
		Mississippian	Horton Group	MHs	conglomerate, arkose, sandstone
		- disconformable to gradational contact -			
		Devonian/ Mississippian	Fisset Brook Formation	DMFb	basalt, rhyolite, tuff, red siltstone and sandstone
		- unconformity -			
		Devonian	Salmon Pool granite	Dsp	red biotite-muscovite granite
		- intrusive contact -			
			Margaree granite	Dm	pink megacrystic biotite granite
		- intrusive contact -			
				Dl	lineated muscovite leucogranite
		- intrusive contact -			
		Silurian or Devonian		SDg	foliated white biotite-muscovite granite
		- gradational contact -			
		Silurian or Ordovician	Jumping Brook complex	Sjbg	greywacke, quartzite, schist
					- conformable contact -
				Sjbp	pelitic schists and gneisses
					- conformable contact -
				Sjbv	basalt, rhyolite, tuff, gabbro, and metamorphosed equivalents
- unconformity -					
Proterozoic	Hadrynian or Cambrian		Cheticamp granodiorite	Hc	grey to buff biotite granodiorite
	- intrusive contact -				
	Hadrynian?			Hm	monzonite, diorite, hornblendite
	- intrusive contact -				
	Helikian?		Red River complex	Hrra	anorthosite, minor mafic schist
	- tectonic contact -				
	Helikian? and younger		Pleasant Bay complex	Hpb	composite granitic gneiss with augen granite gneiss, lit-par-lit gneiss; minor amounts of units Hc, SDg, Dl and Dsp
					- gradational contact -
				Hpbg	strongly foliated, moderately gneissic biotite augen granodiorite and tonalite
					- gradational contact -
				Hpba	gneissic to lineated amphibolite, intercalated with up to 50% of unit Hpb
					- gradational contact -
				Hpbp	biotite gneiss and schist, intercalated with up to 65% of unit Hpb

* Some age designators may differ slightly from those used in Figure 2.

although Jamieson and Craw (1983) attacked this date as unreliable. However gneisses probably correlative to the Pleasant Bay complex gave a Rb-Sr isochron age of 550 ± 80 Ma in the Black Brook area, 25 km east of the mapped area, interpreted to be the age of metamorphism (Cormier, 1972). Rocks lithologically similar to a gabbro-diorite-hornblendite pluton which cuts the gneisses north of Grande Anse River have recently been dated by Rb-Sr isochron at 750 ± 65 Ma in the Baddeck Lakes area, 25 km to the southeast of the mapped area (Jamieson and Craw, 1983). These data suggest that the oldest parts of the Pleasant Bay complex are certainly Precambrian and possibly older than 750 Ma. The youngest parts of the complex may be of Devonian age (Margaree and Salmon Pool granites).

Correlation

Rocks similar to the Pleasant Bay complex underlie much of the central Cape Breton Highlands. Neale (1963) and Milligan (1970) correlated the paragneisses and amphibolites to the George River Group of southeastern Cape Breton Island, but since they also included the younger Jumping Brook complex in this unit, the correlation cannot be completely valid. Wiebe (1972), mapping east of Chéticamp area, distinguished between an older sequence of metamorphic rocks containing crystalline limestone which he termed "George River(?)" and a younger sequence rich in metamorphosed volcanic rocks which he termed "Fourchu(?)". The geology described by Wiebe (1972) closely parallels that observed in the Chéticamp area, except for the rarity

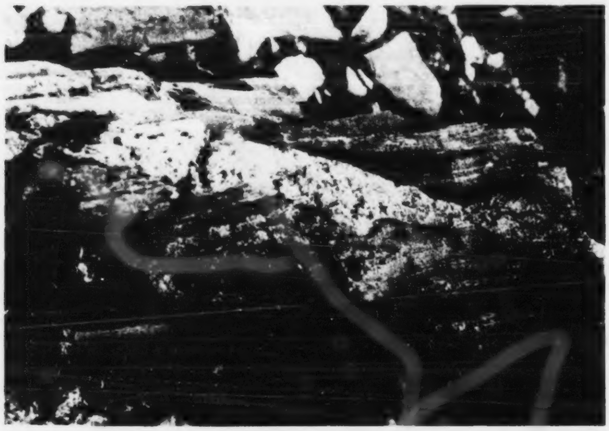


Figure 3. Dyke of Cheticamp pluton cutting granitoid gneiss of the Pleasant Bay complex. Note offset of gneissosity at left centre and slight boudinage of dyke on right side. Corney Brook 3 km from mouth. (GSC 188582)

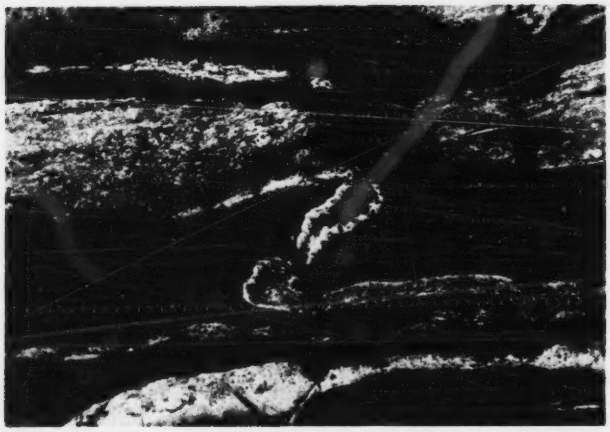


Figure 4. Granitoid dyke ptygmatically folded with host amphibolite. Note leucocratic dyke rim due to later remobilization. Granitoid bodies below hammer handle have been completely transposed into the gneissosity. Mouth of Fishing Cove River. (GSC 188599)

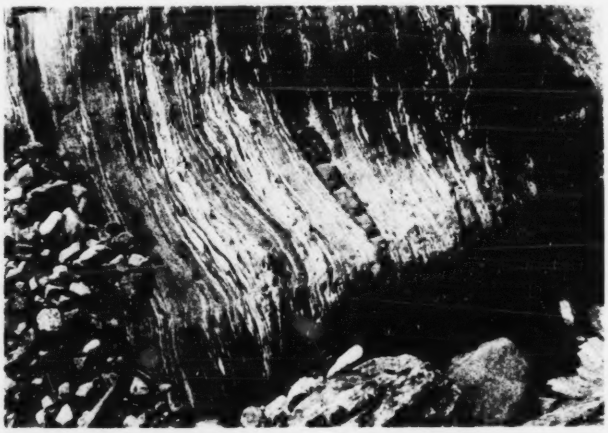


Figure 5. Boudined amphibolite dyke in matrix of biotite augen gneiss. Note the dyke shows neither fine scale banding nor lenticular structure. Chéticamp River about 5 km from mouth. (GSC 188600)

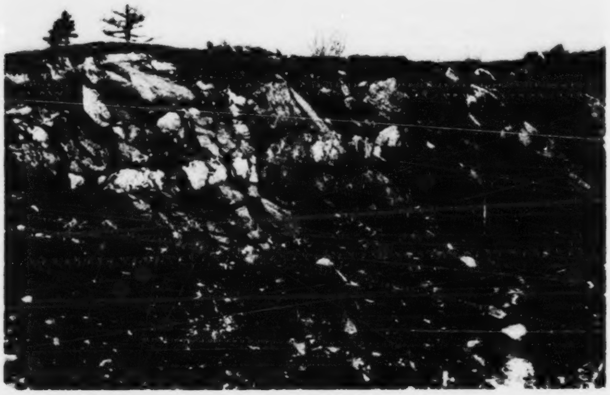


Figure 6. Contact between Pleasant Bay complex (left) and Jumping Brook complex (right). The prominent dyke of Cheticamp pluton is truncated by pale coloured schists (Jumping Brook complex) at the right of the photo, which dip about 45 degrees. The dyke cuts granitoid gneisses of the Pleasant Bay complex which dip more steeply on the left side of the picture, but exhibit a strong schistosity parallel to the Jumping Brook complex along the contact. North side of Jumping Brook about 3 km from its mouth. (GSC 188580)

of marble. Almost identical relations were described in the Cape North region, 25 km to the northeast of the mapped area, by Neale and Kennedy (1975), but their interpretation has been disputed by MacDonald and Smith (1980) who claimed that only one sequence of rocks is present. Jamieson and Craw (1983) took a similar position in the Chéticamp area, arguing that the paragneisses of the Pleasant Bay complex contain only metamorphosed equivalents of the Jumping Brook complex. They agreed, however, that in the Cape Breton Highlands as a whole, both an older "George River Group" and a younger meta-igneous-metasedimentary complex can be distinguished.

Regardless of the disputed presence of basement-cover relations, the term "George River Group" has clearly been applied to rocks of at least two different ages in northern Cape Breton Island, and hence should either be abandoned, or restricted to its type area. Wiebe (1972) pointed out that correlation of rocks from northern Cape Breton Island with type localities of either the George River or Fourchu groups was tenuous at best. The same comment applies even more forcefully to more distant correlations, such as those with southwestern Newfoundland. In these circumstances I have abandoned the traditional "George River" terminology in favour of local names until regional correlations have been more firmly established.

Biotite gneiss and schist *(unit Hpbp)*

Definition and distribution

The Pleasant Bay complex contains enclaves of biotite schist of the Jumping Brook complex (**S**jbp), as well as an older biotite gneiss (unit **H**pbp). Discrimination of older from younger gneisses forms a key to deciphering the Pleasant Bay complex. Older gneisses can be recognized in the field by (a) gneissic, rather than schistose structure, that is by compositional layering, although many of the gneisses tend to part between gneissic layers, (b) presence of potash feldspar porphyroblasts, commonly of two generations, (c) common presence of lits of calcareous (diopside-scapolite) gneiss and marble, and lack of pelitic horizons, (d) presence of dykes and lits of fractured, epidotized, hornblende-bearing granite (Cheticamp pluton). In thin section the older gneisses contain garnet + hornblende commonly with epidote, or more rarely diopside + scapolite in appropriate compositions, whereas younger gneisses contain the lower pressure assemblages anorthite + biotite and muscovite + staurolite with or without andalusite + kyanite.

The older gneisses occur as relatively small enclaves (less than 10 000 m^2) distributed throughout the more granitic foliated rocks (units **H**pb and **H**pbg), from which the high-grade gneisses can be distinguished with difficulty. The foliated granitoid rocks lack potash feldspar porphyroblasts, although plagioclase porphyroblasts or phenocrysts are common and lack strong compositional banding. The contacts between granitoid and metasedimentary gneisses tend to be completely gradational. Old biotitic gneisses can be studied particularly well in the brooks running north into Grande Anse River, along Fishing Cove River, and along the Cabot Trail east of Fishing Cove River.

Lithology

This unit contains a variety of lithologies, some clearly of sedimentary origin, others of more ambiguous character. Marble occurs in three small enclaves on the northeastern margin of the area, forming strongly foliated, colour-banded grey to bluish grey, medium grained carbonate containing lenticular knots of coarse tremolite and diopside up to 5 cm across. Along North Aspy River marble bands 1 to 3 m wide occur within biotite gneisses typical of the rest of unit **H**pbp. Calcareous gneiss, a distinctive pale grey rock containing yellow garnet, as well as quartz lenticles up to 2 cm wide, occurs fairly commonly in this unit, associated with marble along North Aspy River, but intercalated with biotite gneiss south of Grande Anse River, along the Cabot Trail east of Fishing Cove River, and in the canyon of Chéticamp River. This lithology contains coarse grained layers of hornblende + epidote and diopside + scapolite.

The more common biotite gneisses form heterogeneous leucocratic to mesocratic rocks locally reddened by hematite stain. Strong compositional banding varies from 0.5 to 10 cm in width, and the rocks tend to split along gneissosity planes. Gneissic layering may be defined either by variations in the proportions of minerals, with similar textures persisting across layers, or by abrupt alteration of textures due to intercalations of massive, or fractured, or schistose, or foliated layers of diverse composition (compare Fig. 5). As explained above, when the diversity of the layers becomes pronounced, the rock is mapped as composite gneiss (unit **H**pb). The biotite gneiss commonly contains quartzite layers whose texture appears identical to that of the surrounding biotite gneiss, from which it is distinguished only by very high quartz content. Cross cutting dykes are uncommon in this unit, presumably because all such bodies have been transposed into the pervasive gneissic foliation.

In hand specimen, porphyroblasts of feldspar, hornblende and (rarely) garnet may occur, either singly or in some combination. Porphyroblasts commonly show strong alignment, locally in a crenulation lineation. Biotite is ubiquitous but muscovite occurs in this unit only on shear surfaces, commonly associated with chloritic smears. The association of hornblende-biotite in augened porphyroblasts appears characteristic of this unit. Strongly pelitic units appear to be rare or absent. In general the gneisses are relatively coarse grained (2-5 mm) except where cataclastic deformation has produced a fine grained uniform material superficially resembling siltstone.

Contact relations

Intrusive relationships clearly demonstrate that all of the granitic units in the Chéticamp area postdate the biotite gneiss. Similarly at least two ages of metamorphosed basic dykes, distinguished by the presence or absence of potash feldspar porphyroblasts, postdate the biotite gneiss. The relation of the gneiss to the anorthosite of the Red River complex (**H**rra) cannot be unambiguously determined, since the contacts all appear to be intensely sheared. The contact between the older biotite gneiss (**H**pbp) and the younger biotite schist of the Jumping Brook complex (**S**jbp) is exposed, or nearly exposed, east of the Cabot Trail along the lower MacKenzie River and along the upper Fishing Cove River near the bridge on an abandoned part of the Cabot Trail. The contact cannot be precisely located in either case because of minor shearing and presence of abundant granitoid lits, but the relations are similar in both instances. The younger rocks are more uniform, contain disseminated muscovite and no hornblende, do not contain potash feldspar porphyroblasts (although potash feldspar may be present in the matrix), and contain granitic lits only of units **SD**g and **D**l, that is muscovite-bearing granites. The older rocks consistently lack disseminated muscovite, contain hornblende porphyroblasts, and exhibit a complex gneissic structure dominated by diverse granitoid lits and highly deformed lenticles (Fig. 5). With the exception of the locality depicted in Figure 6, the foliation in the two units appears to be parallel, and the contact in some instances appears to be somewhat transitional with a disappearance of feldspar porphyroblasts, and homogenization of the gneiss occurring across as much as 10 m. The vicinity of the contact also contains lenticular enclaves of one type of biotite-rich unit in the other. Such lenticles can only be recognized in areas of unusually good outcrop, and even then definition of the lenticles can be extremely difficult and time consuming because of the similarity of the units. Despite the diffuseness of the contacts, petrographic examination shows a marked divergence in the mineral assemblages of the two units, with pelitic minerals widely disseminated through the younger rocks, and hornblende + epidote through the older. Locally these assemblages are intercalated on a decametre scale, possibly due to strong metamorphism after tectonic juxtaposition. Locally, as along Jumping Brook and on the Cabot Trail south Pleasant Bay, the contrast between the units appears to be enhanced by faulting.

Age

The age of unit Hpbp is unknown, but surely must be Precambrian, and possibly greater than 750 Ma, as previously discussed. The unit may also be composite, in which case the definition of age becomes difficult.

Correlation

This unit includes small occurrences of marble and fairly abundant calcareous gneiss and quartzite layers. Traditionally, any sequence containing marble and quartzite has been correlated to the George River Group of southeastern Cape Breton Island (Milligan, 1970). Such correlations rest on very tenuous foundations, particularly in the present case, because much of the unit contains no marble whatever, few calcareous units (calc-silicates and the like), and very little quartzite. The unit clearly resembles rocks described by Wiebe (1972) in the Ingonish region and rocks described by MacDonald and Smith (1980) in the Cape North region. However the continuing uncertainty concerning the presence or absence of basement-cover relations in northern Cape Breton Island suggests that detailed correlation is premature and imprudent. There may be a basis for correlation northward to Cape North, Saint Pauls Island and southwestern Newfoundland, but this is discussed later.

Gneissic to schistose amphibolite (unit Hpba)

Definition and distribution

The Pleasant Bay complex contains an abundant and easily recognized mafic component, consisting mainly of lineated, amphibolitic gneisses, with lesser amounts of mafic dykes, now commonly transposed into the gneissosity. Outcrops were termed amphibolite if they contained more than 50 per cent amphibolite component, but in many outcrops the proportion of amphibolite approaches 100 per cent. Amphibolitic gneisses tend to constitute all of the outcrop except for variably deformed and transposed granitic dykes (Fig. 4). Amphibolites occur as a minor component in composite gneiss (unit Hpbc) in the form of thin lits, boudins, schlieren and blocks. The term "amphibolite" implies the presence of major amounts of amphibole in the rock, and most examples from the Pleasant Bay complex do contain major amphibole. In some cases the rocks have been sufficiently sheared and/or altered that no amphiboles, or only pseudomorphs, remain. To avoid multiplying the number of units, such mafic chloritic rocks have been grouped with the amphibolites, which they resemble in gneissic structure, mode of occurrence and colour index. The unit is therefore defined as gneisses in which more than half the outcrop is composed of gneiss with a colour index greater than 50. Such rocks occur as irregular, but generally lenticular, enclaves within the gneiss complex. Good examples outcrop along the upper part of Fishing Cove River, in MacKenzie River, and along the Cabot Trail east of the Red River anorthosite complex.

Lithology

The amphibolitic gneiss contains at least three distinct components. One type of material consists of gneisses with banding defined by alternating concentrations of amphibole and feldspar about 1 to 5 cm in thickness. Such rocks commonly contain a little biotite or quartz, and idioblastic garnet may be a prominent constituent of the mafic-rich bands. Amphibole commonly exhibits a moderate to strong lineation. A second type of amphibolitic component exhibits moderate to strong foliation, but relatively weak composition banding, and contains abundant biotite, not uncommonly grown on amphibole, augened potash feldspar porphyroblasts, and, commonly, visible quartz. Such rocks commonly exhibit strong fissility and approach schists rather than gneisses. With decrease in colour index these rocks grade continuously to the biotite gneisses previously described. In some outcrops they form a matrix to belts some metres in width of less fissile, homogeneous gneisses which lack potash feldspar and appear to be younger, since they have clearly not undergone the potash metasomatism which produced potash feldspar porphyroblasts and biotite rims on the amphiboles of the more schistose component. Yet a third component of foliated, but not gneissic, amphibolite occurs in some outcrops, apparently as transposed dykes (Fig. 5).

Occurrences of amphibolite gneiss tend to be small, commonly less than 500 m across strike. Granitoid rocks invariably occur within the amphibolites, occasionally as folded dykes (Fig. 4), but more commonly as schliers, lits or irregular bodies which contain concentrations of the minerals occurring in the surrounding gneiss (amphibole, garnet) and appear to be derived from their surroundings, presumably by partial melting or the extraction of a solute component. The edges, and occasionally the cores, of the amphibolite occurrences tend to be strongly fissile and chloritized, particularly those northeast of Grande Anse River.

Contact relations

The contact relations of the Pleasant Bay complex, discussed above in general terms, apply equally to the amphibolites. The edges of the amphibolite bodies tend to be rich in biotite, and contain potash feldspar and quartz. They grade almost imperceptibly toward the biotite gneisses of unit Hpbp. The relative ages of these units are therefore uncertain, although some of the components of the amphibolites appear to be clearly younger than some of the components of the biotite gneiss (Fig. 5). Foliated but unbanded amphibolites are considered to be correlative to the Jumping Brook complex (unit Sjbv). Such presumed dykes occur commonly within the amphibolites, but no large scale contacts between amphibolitic gneiss and Jumping Brook complex have been observed. North of Grande Anse River a cataclastically deformed, but unfoliated, mafic complex (unit Hm) truncates both mafic gneisses and anorthosite of unit Hrra. The age of this mafic complex is not definitely known, but it slightly resembles the Baddeck Lakes complex dated by Jamieson and Doucet (1983) at 750 ± 65 Ma. The relations between the amphibolite gneiss and the Red River anorthosite complex, although reasonably well exposed along the Cabot Trail and the valley of Red River, remain ambiguous because of intense shearing.

Age

The "age" of the amphibolitic gneiss like that of the biotite gneiss needs to be carefully defined. The oldest component presumably predates the emplacement of the mafic complex noted above, and may therefore be older than 750 Ma. Parts of the complex appear to have undergone fairly intense potash metasomatism prior to development of the pods of granitic rocks which cut across the metasomatized gneiss. This incident of metasomatism and probable partial melting may be connected with emplacement of the Cheticamp pluton, which gives an age about concordant with the metamorphic age of the gneisses (Cormier, 1972), as discussed above. Finally, the foliated but non-gneissic component, if it is correctly correlated with the Jumping Brook complex, presumably is of Ordovician or Silurian age, implying that the homogeneous amphibolitic gneisses which are not metasomatized probably have ages between that of the Cheticamp pluton and that of the Jumping Brook complex.

Correlation

Amphibolitic gneisses in the Cape Breton Highlands were correlated by Neale (1963) and Milligan (1970) to the George River Group, and by Wiebe (1972) to the Fourchu Group. Since the gneisses clearly contain several components of differing ages, and are moderately to strongly metasomatized in many instances, such correlations cannot be correct for all components of the gneiss, and are highly uncertain even for selected components of the gneiss. I have observed similar complex mafic components in the "Coastal gneisses" described by Wiebe (1972) 30 km east of the mapped area, but virtually all high grade gneiss complexes contain one or more mafic components. In the absence of detailed petrographic and chemical data, as well as careful distinction of the various components, reliable correlations cannot be made.

Gneissic biotite augen granodiorite (unit Hpbg)

Definition and distribution

A polychronous, polydeformed sequence of generally granitoid gneisses underlies almost 50 per cent of the area examined. These generally granitoid rocks fall into two distinct subdivisions, according to the amount of non-granitoid material present. The gneissic augen biotite granodiorite (unit Hpbg) is defined by its essentially homogeneous character, strong foliation, relatively weak compositional banding, and consistent presence of feldspar augen. Pegmatitic schlieren together with lits and boudins of more mafic, but strongly assimilated, gneisses occur throughout the unit, but nowhere compose as much as 30 per cent of the volume.

This unit forms a well defined belt extending south from the Grande Anse River beyond the southern edge of the mapped area. This belt is transected by the Margaree pluton (Dm), but lithologies on opposite sides of the pluton appear essentially identical. Much of the unit lies in remote and poorly exposed regions, but very good and typical exposures occur along a foot trail next to McIntosh Brook.

Lithology

Along McIntosh Brook the rocks typically consist of pale grey gneiss, which on closer examination consists of a uniform, medium grained, strongly foliated matrix of biotite, quartz and feldspar, with disseminated augen of plagioclase (Fig. 7). The strongly aligned augen tend to segretate along discrete zones, giving the rock a distinct, but rather weak compositional banding. Augen commonly exhibit not only rounding, but cuspate indentations, or transverse cracks and gaps, suggesting that they have been deformed after initial formation (Fig. 7). Locally the augen coalesce to form zones up to 10 cm wide composed almost entirely of coarse feldspar, but this phenomemon appears to be restricted to the edges of the unit. Pegmatitic schlieren exhibit abruptly gradational boundaries with the gneiss, and may contain wispy, nebulous aligned inclusions of gneiss, suggesting that the schlieren have developed in situ. Mafic schliers and boudins locally lie across the gneissosity, but their boundaries are so nebulous and gradational that the significance of this cross-cutting relation cannot be determined.

Rocks south and east of the Margaree pluton are both poorly exposed and difficult of access. On the headwaters of North Aspy River and on Chéticamp River near the junction with Artemise Brook the rocks are broadly similar to those on McIntosh Brook, but substantially richer in plagioclase and poorer in quartz. In particular they are essentially homogeneous, though strongly foliated, and contain large feldspar crystals in a finer grained, homogeneous, foliated matrix. Locally on North Aspy River the gneisses have been cataclastically deformed to mylonitic gneiss.

Contact relations

West of McIntosh Brook the foliated augen granodiorite appears to truncate biotite gneiss (unit Hpbp). The actual contact could not be found on the highland, but the granodiorite unit strikes north across the trend of the east-northeast gneissosity in the biotite gneiss. The margins of the biotite augen granodiorite appear to display more compositional banding than the core, and the passage from the homogeneous augen granodiorite to the compositionally similar biotite gneiss may be transitional in detail. No large scale relations were observed between units Hpbg and Hpba, but the granodiorite includes relicts of mafic gneiss rich in biotite and containing potash feldspar augen which appear to be derived from the biotitic augen schists of unit Hpba, although the degree of assimilation of the inclusions makes any correlation dubious.

Dykes of foliated but not gneissic amphibolite, thought to be correlative to the Jumping Brook complex, occur within unit Hpbg along the upper part of the North Aspy River. Although they lie generally in the foliation of the granitoid gneiss, and are overprinted by this foliation, in detail they transect compositional boundaries in the host gneiss, and therefore appear to be younger. A major dyke lithologically correlative to the Cheticamp pluton cuts the foliation in the upper part of Mackenzie River. This dyke can be traced almost 3 km, and consists of massive, faintly epidotized, hornblende-biotite granodiorite which cleanly cuts the

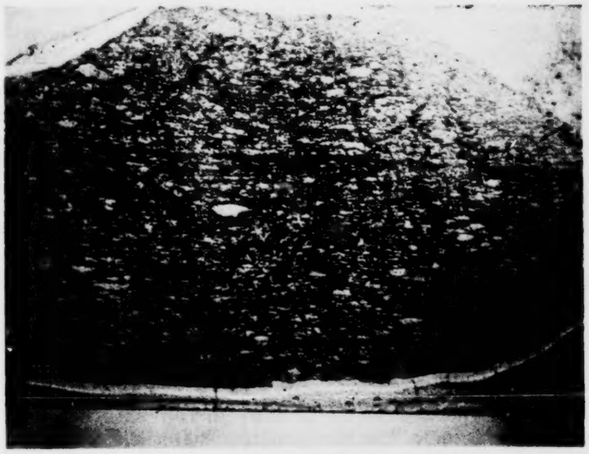


Figure 7. Texture of granitoid gneiss of the Pleasant Bay complex. The prominent lenticles (augen) are composed of polycrystalline quartz and plagioclase. Note cracking of the large lenticle at the top centre, and decussate texture of many lenticles. The dark matrix is biotite. Width of field 3 cm. (Specimen 107, GSC 202868-S)

foliation, but displays neither marginal fining of grain nor contact metamorphic aureole. Characteristic crosscutting porphyritic dykes of the Margaree pluton cut the gneiss south of Mackenzie River near the margins of the Margaree pluton. Locally the gneiss is markedly reddened along these dykes, and exhibits small open fractures perpendicular to the contacts.

Age

According to cross-cutting relations, the gneissic biotite augen granodiorite must be older than the Cheticamp pluton (530 ± 44 Ma, Cormier, 1972), but younger than the biotitic and amphibolitic gneisses (units **Hpbp** and **Hpba**). In general, a late Precambrian age seems probable, but the exact age cannot be specified. As pointed out by Jamieson and Craw (1983) this unit may be composite like the older gneisses, and contain deformed granitoid plutons of several ages.

Correlation

The foliated to gneissic granitic rocks continue south of the study area in a belt depicted by Jamieson and Craw (1983). This belt of deformed granitic rocks extends from Grande Anse River almost 75 km to the vicinity of Lake Ainslie, and forms one of the major features of the geology of the Cape Breton Highlands. It may represent one or more elongate igneous complexes, analogous to, and parallel with, the younger Margaree pluton. The layers of foliated granite within the surrounding gneisses could be, in part, dykes from this massif, although it seems probable that many of these layers represent anatectic melt. Olszewski et al. (1980, 1981) have recently documented the presence of gneissic granitoid rocks of 700-800 Ma age southeast of the Cape Breton Highlands and in other parts of the Canadian Appalachians. Although correlation cannot be confidently made, such rocks form possible correlatives of the gneissic granitoid rocks of the highlands.

Lit-par-lit and composite granitoid gneisses (unit Hpb)

Definition and distribution

After separation of other units of the Pleasant Bay complex, a large number of outcrops remain unassigned, namely those with a sufficiently hybrid character that they do not fit into other units. These left-overs have been gathered together as "composite gneiss," a term introduced by Neale (1963). Despite the residual nature of this unit, it can be defined reasonably satisfactorily for mapping purposes. Rocks of this unit exhibit a gneissic granitoid matrix containing more than 20 per cent by volume of non-granitoid material. Such rocks can be examined at numerous localities within the mapped area, particularly along the Cabot Trail from the head of Jumping Brook to Mackenzie River. In broad terms the unit forms a north-south belt extending from Mackenzie River south to the southern limit of the mapped area. A narrower, parallel belt can be examined in the upper parts of Corney and Robert brooks and on Chéticamp River.

Lithology

These gneisses can be divided into two types, depending on the character of the non-granitoid material. Where this material occurs in parallel layers, the rocks form lit-par-lit gneiss. Where the material has an irregular form the gneisses have an agmatitic character. Lit-par-lit gneisses commonly include at least four lithologies, namely schistose biotite gneiss (unit **Hpbp**?), highly chloritized, generally fissile mafic rocks (unit **Hpba**?), foliated biotite granite to granodiorite (unit Hpbg?), and a massive, faintly epidotized granitic component (unit H̲c?). Pure quartz bands, presumably transposed quartz veins, occur locally. Commonly the bands range from 1 to 100 cm in width, although thicker bands occur locally. Each band exhibits sharp, parallel contacts with its neighbours. Crosscutting dykes occur rarely (Fig. 8), but lenticular, partially cross-cutting masses of leucocratic muscovite granite, commonly hematite-stained, occur fairly commonly, particularly near the margins of the unit. Individual layers, particularly mafic ones, may be intensely sheared, with all mafic minerals converted to chlorite. Shear surfaces commonly have hematite coatings. Some fine grained, uniform biotite-gneiss layers may be recrystallized mylonites, since they are much finer grained and more uniform than the surrounding biotite gneisses which they compositionally resemble.

The other type of composite gneiss contains non-granitoid material as blocks or schlieren, commonly elongate along the gneissosity. The size of the blocks varies from a few centimetres to 20 m in length, or possibly more since such large blocks are rarely completely exposed. The shapes of the blocks range from angular, suggesting rigid fracture of larger masses, through various degrees of rounding to the most common spindle-like shapes. The blocks consist of biotite gneiss and amphibolite, now strongly overprinted by potash feldspar augen, and by biotite overgrowing amphibole. Where gneissosity can be read in the blocks, it everywhere parallels that in the matrix, but many of the blocks exhibit no foliation. The matrix consists of a leucocratic granitoid phase, commonly with a strongly lenticular texture. Mafic minerals tend to be almost entirely confined to narrow lenticles. Locally the texture may be pegmatitic, but all gradations occur towards a finer grained, pale grey aplitic texture typical of unit **SDg** to which the composite gneiss appears to grade.

Contact relations

Few diagnostic contact relations have been observed between the lit-par-lit gneisses and the other plutonic units. Dykes thought to be referable to the Salmon Pool granite cleanly cut the lit-par-lit gneiss (Fig. 8). Lenticular masses of leucocratic muscovite granite (unit **Dl**) locally crosscut, but are generally conformable to the foliation. Layers of granitoid rocks similar to the Cheticamp pluton form lits within the gneiss. In upper Robert Brook unfoliated granodiorite of the Cheticamp pluton abruptly truncates the

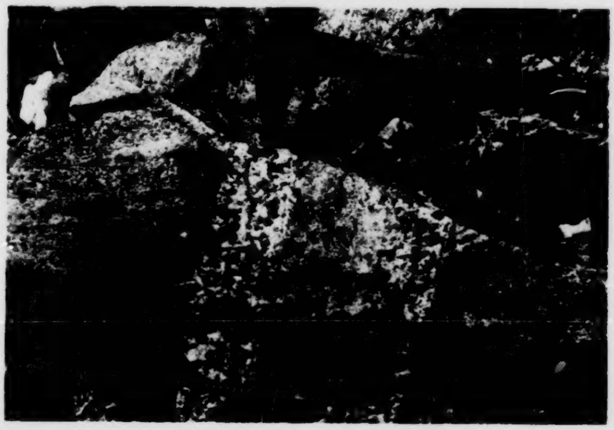


Figure 8. Dykes of massive Salmon Pool granite cutting composite gneiss of the Pleasant Bay complex. Cabot Trail east of Fishing Cove River. (GSC 188586)

composite gneiss. The Cheticamp pluton contains few, small, angular basic inclusions whereas inclusions in the composite gneiss to the east range up to several metres across, and have very sinuous shapes. The gneissosity of the matrix is strongly contorted along the contact. A few hundred metres farther up the brook, equivocal relations between composite gneiss and Jumping Brook complex are exposed in a series of cliffs. The composite gneiss gradually loses its inclusions toward the contact, and the matrix becomes finer grained, more homogeneous, and muscovite-bearing, so that the phases actually in contact are a pale grey aplitic gneiss and an amphibole schist. The latter shows no penetration by the former, and hence must be younger than the amphibolite blocks within the composite gneiss. Similar relations occur on the Cabot Trail on the rise from Mackenzie River south to the highlands, although in this case the transition is from composite gneiss, through aplitic gneiss to biotite-muscovite gneiss of the Jumping Brook complex. These relations suggest that the composite gneiss originally formed before the Jumping Brook complex, but has subsequently been reactivated, welding together the Jumping Brook and Pleasant Bay complexes.

Age

The completely concordant character of the diverse lithologies in the lit-par-lit gneiss suggests that the final configuration of the gneiss postdates the formation of the granitic portion. MacLaren (1955) referred to the rocks as "injection gneiss", but large scale injection of magma could not achieve so uniform a banding without local boudinage and cross-cutting relations. It seems more likely that parallelism was achieved by structural means after, or during, assembly of the diverse components, rotating any previously cross-cutting elements into the foliation. On this view final formation of the lit-par-lit gneiss accompanied or postdated the Cheticamp pluton, elements of which appear in the gneiss, but predated the Devonian granitoid rocks, which cut the gneissosity.

Some of the components of the composite gneiss clearly must be pre-Cheticamp in age, since they are cut by the pluton. On the Cabot Trail at the bridge over Mackenzie River the rocks resemble agmatitic composite gneiss, but the matrix is unusually homogeneous, fine grained, and contains muscovite and fluorite. The radiometric ages obtained from this location are 367 Ma (Rb/Sr biotite age), 413 Ma (Rb/Sr muscovite age) both by Fairbairn et al. (1960), and 364 ± 35 Ma (two point Rb/Sr isochron) by Cormier (1972). Wanless et al. (1968) obtained similar ages a few kilometres to the south. These Devonian ages fix the age of latest isotopic homogenization, presumably resulting from remobilization, but they do not require a Devonian age for the protolith, and such an age appears highly improbable, as previously discussed. I interpret the composite gneiss to be Precambrian, but partially remobilized during the Devonian giving rise to gneiss of unit SDg.

Correlation

Jamieson and Doucet (1983) described an "Egypt highland" unit 25 km south of the mapped area which consisted of schistose to gneissic granitic rocks, cut by numerous zones of very fine grained micaceous schists, which they suggested might be small shear zones. The presence of numerous small shear zones suggests similarities to the lit-par-lit gneiss of the Chéticamp region. The Coastal gneisses of Wiebe (1972) appear to be similar to the agmatitic composite gneiss with blocks of assimilated mafic material. Neale (1963) described presumably correlative gneisses as extending northeast and east of the mapped area. Detailed correlation of this unit cannot be made at present due to lack of detailed information, but correlatives are probably present wherever correlatives of the Pleasant Bay complex occur.

Red River anorthosite complex *(unit Hrra)*

Definition and distribution

Neale (1963) discovered the presence of anorthosite in the Cape Breton Highlands and mapped several small masses. The largest, termed the Red River anorthosite by Jenness (1966) and the Red River anorthosite complex by MacDonald and Smith (1979), outcrops in the northeast corner of the mapped area, and a small, presumably related lenticle outcrops in a creek valley 3 km farther to the south. The latter occurrence near the headwaters of Grande Anse River is the southernmost known occurrence of anorthosite in Cape Breton Island. The Red River anorthosite complex forms a lenticular mass about 10 km long by 1 km wide, the southern third of which lies within the mapped area. Outcrop occurs only along roadcuts on the Cabot Trail, in the valley of Red River, and in an unnamed creek valley tributary to the Red River. The map unit consists of white to pale grey anorthosite and rather minor amounts of anorthositic rocks with colour indices greater than 10.

Lithology

The anorthosite commonly exhibits a white to pale grey, rather lumpy outcrop apearance. On fresh surface a pale green tinge, or a network of pale green streaks due to epidotization is characteristic. Virtually all outcrops exhibit an intricate network of fracture and shear zones, and the rocks tend to break on complex hackly surfaces, indicating that it forms a slightly re-cemented monomict breccia. The colour index commonly lies below 10, possibly due to destruction of some mafic minerals by brecciation and alteration. Rare clots of chloritized hornblende form the most common mafic minerals. Occasionally these clots show a weak foliation, but the anorthosite rarely shows strong foliation except in well defined shear zones. Along the margins of the complex such zones contain mesocratic gneisses in which anorthosite layers up to several centimetres thick form a component. These gneisses exhibit severe shearing between compositional layers, and brecciation within layers, suggesting that they are tectonically assembled sequences of slices. MacDonald and Smith (1979) reported a sliver of pyroxene granulite facies metamorphic rocks associated with the complex, but this outcrop was not observed during the present study.

Contact relations

The eastern and western contacts of the Red River anorthosite complex lie within major shear zones several hundred metres wide. Many apparent layers or lenticles of the anorthosite in these zones are probably tectonic slivers. Relations between the anorthosite and the gneisses to the east and west remain uncertain due to these tectonic contacts. The anorthosite seems clearly intruded by a younger gabbro-hornblendite-diorite complex (unit $\underline{H}$m) which crosscuts the older gneisses (unit Hpbp) north of the Cabot Trail along Grande Anse River. Although this complex is itself severely sheared and shattered in the shear zones, dykes of similar material cleanly cut the anorthosite. Along the Cabot Trail a few dykes of massive, pale coloured aplitic granite, slightly shattered and altered, cut the anorthosite. These dykes resemble the Salmon Pool granite (unit Dsp).

Age

No well founded estimate of age can be given for the Red River anorthosite complex since its contact relations are so poorly known. Many North American and European anorthosites of this type have emplacement ages of 1400 to 1800 Ma (Emslie, 1978), which might be taken as a preliminary "guesstimate" of the age of the Cape Breton anorthosite.

Correlation

Correct correlation of the Red River complex forms one of the keys to deciphering the broad scale structure of northern Cape Breton Island. The anorthosite clearly forms fragments exotic in their present surroundings, a point emphasized by the fragment of pyroxene granulite facies metamorphic rocks (MacDonald and Smith, 1979), a grade appropriate to anorthosite complexes but quite different from that of the surroundings. Two possibilities exist, namely that the anorthosite was emplaced primarily by lateral transport, or alternately that it was emplaced primarily by vertical transport. If the anorthosite was emplaced primarily by horizontal transport, the mylonites and shear zones bounding the anorthosite must be zones of major transcurrent movement, and discrepancies would be expected between the geology on opposite sides. This appears not to be the case. Neale (1963) showed marble, quartzite and distinctive reddish brown monzonites on both sides of the anorthosite. It seems more likely that the anorthosites have been exposed mainly by vertical movements, and form a horst of basement in the Pleasant Bay complex, as advocated by MacDonald and Smith (1979). Large scale uplift of basement would be expected to produce a positive gravity anomaly, and the regional gravity map does show a gravity high along the trend of the anorthosite belt (Earth Physics Branch, 1979).

White or pale grey anorthosites are typical of highly deformed anorthosite complexes in the Grenville province of Quebec, Ontario and New York, the southern Long Range of Newfoundland, and western Norway. The nearest well-known anorthosite masses of similar character lie in the Long Range of southwestern Newfoundland (Herd and Dunning, 1979). Although anorthosite is a relatively rare rock, it can hardly be presumed that two occurrences separated by such a distance are directly correlative. The distribution of anorthosite, however, provides data for plate tectonic modelling which is discussed later under Tectonic Setting.

Monzonite, diorite, hornblendite *(unit Hm)*

Definition and distribution

The Red River and Pleasant Bay complexes contain a characteristic suite of hornblende-dominant younger intrusions ranging from hornblendite to diorite to monzonite in composition. The largest and most accessible member of this suite outcrops in the northeastern corner of the mapped area, but similar bodies too small to show on the map occur at the head of Jumping Brook, between the Cabot Trail and North Aspy River, and south of Chéticamp River at the head of the LeBlanc Brook. The intrusions tend to be massive in the central parts and badly fractured to weakly foliated at the margins. Material in the Grande Anse River valley has been sheared to chlorite paste, but elsewhere the grain size ranges from coarse to pegmatitic, and amphibole predominates over other mafic minerals. The unit can be defined by its hornblende-dominant character, since no other massive igneous rocks in this region have hornblende as the dominant mafic.

Lithology

Rocks of this unit vary from coarse hornblendite through hornblende-plagioclase rocks, locally pyroxene-bearing (termed diorite by Neale, 1963). Except where sheared, all have grain sizes exceeding 2 mm, and the grain size commonly exceeds 1 cm. Parts of the more mafic rocks consist essentially of hornblendite in which the amphibole is visible poikilitic with plagioclase. Much of the western and southern parts of the mass north of Grande Anse River has been reduced to a homogeneous chloritic paste, commonly hematite-stained. These rocks can be distinguished only with difficulty from sheared amphibolites of the Pleasant Bay complex. More massive lenticles, however, occur within the complex which have massive, euhedral to subhedral amphibole fabrics, quite unlike the amphibolites of the Pleasant Bay complex. Despite the ultramafic character of some of the rocks (hornblende greater than 90%) rare grains of quartz can commonly be identified with a hand lens. The plagioclase of this unit commonly has a brownish shade, suggestive of alteration. Along the Red River a small body of this hornblende-plagioclase rock shows good mineral layering which appears to be a primary igneous phenomenon.

Contact relations

A near-contact with anorthosite occurs in a stream flowing into Red River. The anorthosite appears markedly epidotized and badly shattered. Diorite, 5 m away, is essentially massive though severely altered. The presence of dykes of this hornblendic unit in the anorthosite was noted above. Mafic rocks clearly truncate the Pleasant Bay complex south of Red River. Relations with the Cheticamp pluton remain uncertain. The two units resemble each other in general degree of preservation, notably the generally massive, but cataclastic and altered character. On Mackenzie River, an outcrop of Cheticamp-type granodiorite appears to be a dyke cutting massive hornblende-plagioclase rock but the contacts are not visible. The mafic rocks along the Cabot Trail and north of Grande Anse River are cut by fine grained biotite granite dykes, thought to be referable to the Salmon Pool granite, and by quartz-feldspar porphyry dykes probably referable to the Margaree granite.

Age

The hornblendite-diorite unit clearly postdates the anorthosite and predates the Devonian granitoid rocks. The relations with the Cheticamp pluton remain unclear, but the Cheticamp pluton is probably younger. The best estimate of the age appears to lie between the 530 Ma age of the Cheticamp pluton (Cormier, 1972) and the 1400+ Ma "guesstimate" for the anorthosite.

Correlation

No correlatives for the hornblendite-diorite suite are presently known to the south or east of the mapped area, unless the Baddeck Lakes complex (Jamieson and Doucet, 1983) may be correlative.

Cheticamp pluton *(unit Hc)*

Definition and distribution

The Cheticamp pluton (formerly termed the Corney Brook complex by Currie, 1975) forms an elongate mass of granitoid rocks extending south from Jumping Brook beyond the mapped area. According to the compilation of Jamieson and Craw (1983) the total length of this pluton exceeds 25 km, although the width nowhere exceeds 5 km. The Cheticamp pluton can be examined along the Cabot Trail in cliffs extending from Pillar Rock to Jerome Brook, and in Jumping Brook, Corney Brook, Robert Brook, Fiset Brook and Farm Brook.

The Cheticamp pluton consists of rather coarse, variably brecciated and chloritized, biotite (± hornblende) granodiorite. Although the pluton nowhere exhibits foliation, deformation appears strong, and along Corney Brook the rocks grade from brecciated granodiorite to cataclastic rocks whose granitoid protolith is not obvious (unit Mbr). Outcrops were assigned to the Cheticamp pluton if they were unfoliated granitoid rocks displaying healed brecciation, or other evidence of internal disruption, and alteration of mafic minerals. An additional useful, but less common, criterion is the presence of ivory-coloured spots or patches due to clumps of faintly epitoditized plagioclase grains. Rocks of this description not only form the major pluton defined above, but occur as a major suite of dykes and lits within the Pleasant Bay complex, particularly in the region stretching from the head of Jumping Brook to the head of McIntosh Brook, that is the region roughly along trend of the major pluton. A few of these satellite masses are large enough to appear in Figure 2, but most form an unmappable component of the composite gneiss (unit Hpb).

Jamieson and Craw (1983) described the Cheticamp pluton south of the mapped area as undeformed and locally muscovite-bearing. The definition for the unit given above presumably would not be adequate in this region, but it works well within the mapped area. Even on Farm Brook near the southern boundary of the area, the granitoid rocks, though less disturbed than farther north, contain numerous breccia zones, are altered by chlorite and hematite, and have been moderately reactivated, as indicated by presence of boudined and assimilated dykes.

Lithology

The Cheticamp pluton consists of variably brecciated and altered biotite granodiorite grading to diorite. The outcrop surface typically has a rough, uneven surface which may display fragments (Fig. 9) or exhibit an intricate pattern of planar and curved fractures (Fig. 10). Breccia veinlets occur in many outcrops, and the whole outcrop may be composed of brecciated, or fractured and recemented rock. Quartz and feldspar commonly form pods or polygonal areas up to several centimetres across surrounded by a network of biotite and greenish chlorite. In many specimens chlorite is pseudomorphous after hornblende, and unaltered hornblende occurs in a few specimens. The rocks vary from leucocratic to mesocratic, but most specimens appear dingy green because of ubiquitous chlorite alteration. Cognate dykes and lenses of more leucocratic material abound near the margins of the unit, and quartz veins are locally abundant in these zones. These vein and dyke complexes outcrop in Jumping, Robert, Fiset and Farm brooks. On Farm Brook the western part of the pluton has been overprinted by microcline megacrysts of rounded to euhedral shape up to 5 cm across. This rock strikingly resembles fractured parts of the Maragree pluton along North Aspy River.

Contact relations

As previously discussed, the Cheticamp pluton clearly intrudes the Pleasant Bay complex, and probably intrudes amphibolite of unit Hm in the form of dykes. Relations between the Cheticamp pluton and unit Hpb of the Pleasant Bay complex can be studied in some detail in Robert Brook where more massive and homogeneous fractured granodiorite of the pluton truncates agmatitic gneiss of the Pleasant Bay complex. Similar, but poorly exposed, relations occur along the north side of the Chéticamp River east of the mouth of Robert Brook.

Relations between Cheticamp pluton and Jumping Brook complex can be examined on Jumping Brook, east of Pillar Rock and on Faribault Brook. At all these localities the Jumping Brook complex appears to be younger. The contact is reasonably sharp (defined within 5 m). No Cheticamp dykes cut the Jumping Brook complex, nor are there discernable contact effects. Dykes of Cheticamp type are locally truncated by schists of the Jumping Brook complex (Fig. 6). On Jumping Brook and east of Pillar Rock the contact has clearly been affected by shearing, and on Faribault Brook about 15 m is covered at the contact. It might be argued that the above relations are structural effects which do not define relative age. Nevertheless the

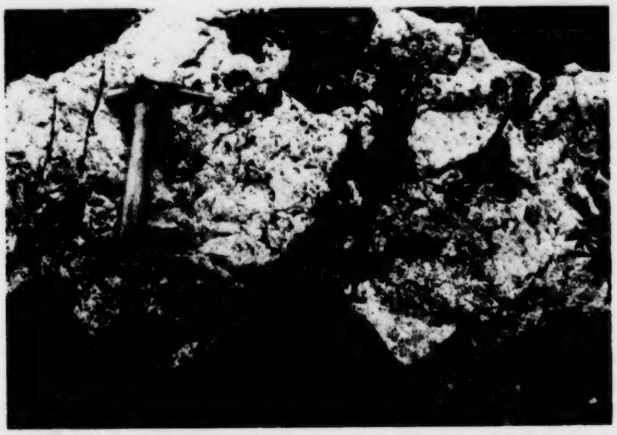


Figure 9. Cheticamp pluton, mouth of Corney Brook. Note the irregular surface of the rock which consists of fist-sized fragments welded together. (GSC 188596)

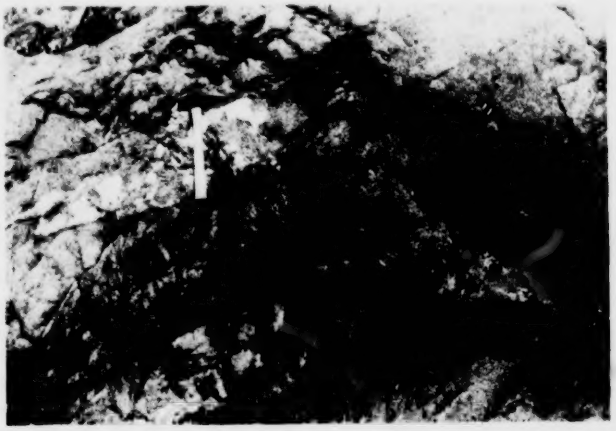


Figure 10. Heavily fractured Cheticamp pluton cut by an undeformed dyke of muscovite leucogranite under the hammer. Note the curving fractures, and the anastamosing network of fractures in the Cheticamp pluton. Corney Brook, 1 km from mouth. (GSC 188569)

age relations outlined above are supported by dyke relations within the Cheticamp pluton, particularly on Fiset and Farm Brooks and east of Pillar Rock. Mafic dykes of two types occur. Fresh, locally vesicular dykes with chilled margins are either unaltered, or fractured and chloritized. Such dykes strongly resemble the Devonian-Mississippian Fisset Brook Formation with which they are here correlated. The other dykes exhibit foliation but not gneissosity, tend to be slightly sinuous in plan, and are locally boudined. They have clearly been affected by subsequent remobilization of their host. These dykes are correlated with the Jumping Brook complex, and indicate that the Cheticamp pluton is older than the Jumping Brook complex. Even older, nebulous mafic inclusions occur locally within the Cheticamp pluton, presumably derived from the Pleasant Bay complex.

Age

Cormier (1972) obtained a 10 point Rb-Sr isochron age of 530 ± .44 Ma with an initial $^{87}Sr/^{86}Sr$ of 0.707 from samples supplied by D.G. Kelley. Jamieson and Craw (1983) attacked this age as unreliable because of dubious sampling. Cormier (1972), however, specified that the samples came from beneath the Fisset Brook Formation which, according to both Jamieson and Craw (1983) and the present mapping, would require that the samples came from the Cheticamp pluton. A new and undisputed sample of the Cheticamp pluton has been collected (R.A. Jamieson, personal communication, 1982) from which zircons will be concentrated and dated. This may resolve the present ambiguity over the age. At any rate the pluton must be older than the 440 ± 7 Ma zircon concordia age obtained from granitic dykes cutting the pluton east of Pillar Rock (Currie et al., 1982).

Correlation

Few plutons in the Cape Breton Highlands have yielded ages similar to that of the Cheticamp pluton, but such ages are common in southeastern Cape Breton Island (Keppie and Smith, 1979). There seems no obvious reason why granites of this age should be absent from the highlands, and Olszewski et al. (1981) reported that the southeast corner of the highlands contains a 553 Ma granite cutting basement of about 800 Ma. These relations would be reasonable for the Cheticamp pluton and Pleasant Bay complex respectively. Cormier (1972) reported a 550 Ma Rb-Sr isochron for gneiss 30 km east of the mapped area. This was attributed to isotopic re-equilibration at this time associated with high-grade metamorphism.

Jamieson and Craw (1983) and Barr and MacDonald (1982) implied that the Cheticamp pluton intrudes the Jumping Brook complex, and forms a massive, essentially undeformed pluton. These relations do not hold in the area mapped by me. Barr and MacDonald (1982) found that the southern part of the Cheticamp pluton locally contained muscovite, which also is atypical of the rocks in the mapped area which locally contain hornblende. These relations suggest that two plutons may have been confused. I have briefly examined the supposed continuation of the Cheticamp pluton south of the mapped area. Although many rocks are similar in mineral composition, these rocks are much less deformed than those examined north of Farm Brook. Possibly the differences may be transitional, but there seem to be real differences between the northern and southern extremities of the pluton which need to be explained.

Jumping Brook complex

Definition and distribution

The Jumping Brook complex is defined to embrace all metamorphic rocks in the region which exhibit schistosity, but not gneissosity, that is all rocks with strong foliation and/or lineation which lack small scale compositional banding and are not granitoid. Such rocks outcrop in a discontinuous south-trending belt extending from north of Jumping Brook to beyond the south edge of the mapped area. In addition to these large areas of schist, smaller enclaves occur within the Pleasant Bay complex, and many schistose amphibolitic dykes occur within the Cheticamp pluton. Fine grained schistose felsite or microgranite dykes occurring on cliffs east of Pillar Rock have likewise been included within the Jumping Brook complex, since they appear to feed rhyolite in the overlying Jumping Brook complex. The complex is named from Jumping Brook, along which a good section is exposed.

The definition of the Jumping Brook complex as schistose but not gneissic, and the Pleasant Bay complex as gneissic (and possibly also schistose, although most specimens are not), leaves in doubt the assignment of specimens which are strongly schistose, but have a weak or moderate gneissosity. Such rocks are relatively rare, suggesting that the adopted definitions are appropriate mapping criteria, but they do occur from the head of Jumping Brook to Fishing Cove River. These rocks appear to be truly transitional between the two complexes, and possibly indicate some degree of tectonic or metamorphic mixing of the two units.

Lithology

The Jumping Brook complex consists of three units, namely a unit rich in basaltic rocks with minor rhyolitic material, pelitic schist and greywacke (unit Sjbv), a unit which consists essentially of pelitic schists with some interbedded greywacke (unit Sjbp), and a unit consisting of monotonous quartz-rich greywacke (unit Sjbg). The greywacke is relatively mafic and rich in chlorite, suggesting that it contains debris from the basaltic unit. Lenticles of amphibolitic or gabbroic rocks occur throughout the complex. The internal stratigraphy remains uncertain since primary sedimentary structures are nowhere preserved within the complex.

The Jumping Brook complex preserves some spectacular progressive metamorphic sequences ranging from sub-chlorite grade to anatexis. The progression has probably been tectonically telescoped, but essentially complete transitions can be observed in the upper parts of Corney Brook and Trout Brook.

Contact relations

The contact relations of the Jumping Brook complex with older units have previously been summarized. The complex exhibits abrupt transitions with the Pleasant Bay complex, and appears to overlie the Cheticamp pluton, presumably unconformably. Mafic dykes, assigned to the Jumping Brook complex on the criteria listed above cut both the Pleasant Bay complex and the Cheticamp pluton. Along Corney Brook, Robert Brook, and the Cabot Trail south of Mackenzie River, the Jumping Brook complex is bounded by a white or grey uniform fine grained gneissic biotite-muscovite granite (unit SDg). This rock contains wispy enclaves of, and appears to grade to greywacke or semi-pelitic schist of the

Jumping Brook complex. I interpret this granite to have developed by high grade metamorphism or anatexis of part of the Jumping Brook complex. The western part of the complex from Pillar Rock north to Corney Brook has been severely cataclastically deformed, and grades to purely cataclastic rocks (unit **Mbr**). These cataclastic rocks may be bounded by abrupt margins, or the transition to normal schists may be almost imperceptible, so that in this region the significance of foliation must be assessed with caution. Large slabs of the Jumping Brook schists appear in conglomerages of the Horton Group south and east of Presqu'ile. The Salmon Pool granite intrudes the Jumping Brook complex along Chéticamp River near the mouth of Daphiné Brook. The Jumping Brook complex is fractured and reddened at the contact.

Age

A fine grained, foliated granite dyke cutting the Cheticamp pluton east of Pillar Rock, and thought to be a feeder to the rhyolites of the Jumping Brook complex at the top of the cliffs in this locality, returned a concordia intercept age of 439 ± 7 Ma for a U-Pb age on zircon, and a K-Ar age of 369 ± 9 Ma on biotite respectively (Currie et al., 1982). These dates are interpreted to indicate emplacement and metamorphic or cooling ages respectively, and to indicate that the Jumping Brook complex is latest Ordovician or Silurian and was metamorphosed in the Devonian. These age relations are compatible with the contact relations outlined above.

Correlation

Jamieson and Craw (1983) regarded the Jumping Brook complex as equivalent to a metamorphosed volcano-sedimentary complex mapped by Jamieson and Doucet (1983) in the Middle River area, 30 km south of the mapped area, to a sequence mapped by Wiebe (1972) on Clyburn Brook and surrounding areas about 15 km east of the mapped area, and to a sequence in the Cape North region 25 km northeast of the mapped area. If these correlations are accepted, strata correlative to the Jumping Brook complex form a chain of elongated occurrences fringing the central part of the Cape Breton Highlands. Wiebe (1972) and Neale and Kennedy (1975), as well as the present work, argued that the volcano-sedimentary complex was younger than surrounding, more granitoid, gneiss. MacDonald and Smith (1980) and Jamieson and Craw (1983) claimed that the surrounding gneisses represent high grade equivalents of the volcano-sedimentary schists. In part this argument may be semantic, since all parties appear to agree that the schists have locally been strongly metamorphosed and pass gradationally to gneiss. Moreover it is by no means clear that the geology of the Cape Breton Highlands is well enough known that correlation of the several distant segments of metamorphic rocks can be confidently made. Two or more different schist complexes may be involved.

Volcanogenic schist (unit Sjbv)

Definition and distribution

Schists of the Jumping Brook complex have been included in this unit if they are rich in amphibole or chlorite, or if they are granitoid (rhyolitic) in appearance. Strongly chloritic mafic schists predominate in the lower part of Trout Brook, and chemical analyses suggest that they are of volcanic derivation. Material of rhyolitic (mylonitic) appearance occurs along the top of the cliffs east of Pillar Rock, and along the road from there to the garbage incinerator. Chlorite-rich schists of intermediate character abound in the upper parts of Faribault and Daphiné brooks. This material is interlayered with more quartz-rich materials, and may represent metamorphosed tuffs. This zone also contains outcrops of more homogeneous schistose amphibolite, apparently relicts of dykes or small stocks. North of Chéticamp River amphibolite locally bounds the Jumping Brook complex, varying in character from relatively massive material akin to hornblendite, to biotite amphibolite with a granitic fraction.

Lithology

From this summary, it can be seen that no undisputed primary criteria of volcanic origin remain in any of the rocks of the Jumping Brook complex, although phenoclasts appear to be present in some of the rhyolitic rocks. Deduction of a volcanic origin can only be made from their appearance, correlation and their chemical composition. The lower grade rocks in Trout and Faribault brooks exhibit marked compositional layering which appears to be at least part primary, though no doubt transposed and accentuated by the strong schistosity. The rocks are fine grained and homogeneous without prominent partings between layers, although parting planes are elsewhere prominent in the rocks. Chlorite and epidote are disseminated throughout the rock, and in some places fill small tension gashes in the rock along with prehnite. Fine grained intensely fissile, crenulated homogeneous pink schists occur in this unit at Pillar Rock, on Trout Brook and on Faribault Brook. Although presently mylonitic, these rocks lie within the mafic schists and appear to form part of the sequence. They contain augened potash feldspar crystals of a distinctive character which appear to be phenoclasts. The general appearance and disposition of these acid and mafic schists is consistent with an origin from acid and mafic tuffs.

The more homogeneous basic rock in Faribault Brook, mapped as gabbro by MacLaren (1955), consists of intensely schistose, chloritized medium to coarse grained rocks. They clearly seem to be relicts of gabbroic material, but their original form and mineralogy cannot now be discerned due to deformation and poor outcrop. The amphibolites extending north from Chéticamp River exhibit intense schistosity and good mineral lineation of hornblende. Alteration is not obvious and the rocks are homogeneous, coarse grained amphibolites. Unlike amphibolites of the Pleasant Bay complex, garnet is rarely visible. Most of these outcrops contain 10 to 20 per cent of granite component, either in the form of schistose conformable enclaves, or more rarely as anastomosing veinlets.

Age

The emplacement and metamorphism ages of 439 ± 7 Ma and 369 ± 9 Ma, previously referred to, are derived from acid rocks of this unit.

Contact relations

Many of the contact relations deduced for the Jumping Brook complex depend upon relations of foliated amphibolite dykes attributed to the Jumping Brook complex. These dykes are identical to the amphibolites marginal to the complex in that they are essentially homogeneous, schistose amphibolites. In the case of dykes, grain size varies from medium grained to coarse grained, whereas the amphibolites within the main body of the Jumping Brook complex are consistently coarse grained.

Correlation

No other volcanic rocks of Late Ordovician to Silurian age are known from Cape Breton Island. The nearest volcanic rocks of equivalent age occur in the Bear Brook

Formation on the northern mainland of Nova Scotia (Keppie, 1979). However volcanic rocks of this age are not rare in the Atlantic provinces, for example the Springdale and Botwood formations of Newfoundland. Long range correlations are clearly premature when the age of the Jumping Brook complex remains contested, but there is no strong geological reason why Silurian volcanic rocks should not be present in northwestern Cape Breton Island.

Pelitic schist *(unit Sjbp)*

Definition

Pelitic rocks of the Jumping Brook complex can commonly be recognized by strong, commonly crenulated, schistosity and abundance of muscovite. Pelitic rocks are interbedded with mafic chlorite schists in Trout Brook and in the upper parts of Faribault and Daphiné brooks, but elsewhere form a layer some tens to hundreds of metres thick which lies closer to the centre of the complex than the mafic schists of unit Sjbv. This relation is compatible with the pelitic rocks overlying the mafic schists, but no direct evidence of such a relation was found. In addition to the localities noted above important sections of pelitic rocks occur in Chéticamp River, Corney Brook, Trout Brook and along the Cabot Trail south of Mackenzie River.

Lithology

At low grade, as in Faribault Brook, the schists have pale to dark greenish tints, or even greenish black colour in Trout Brook. The most abundant pelitic rocks, however, are distinctive silvery schists, often with spectacular knotted or nodular surfaces. Locally such schists have a pale pink tinge due to hematization. Muscovite-rich schists typically occur interbedded with more quartz-rich and less fissile material on a scale of a few centimetres. In many cases porphyroblasts of metamorphic minerals distinctive of grade are confined to thin layers within the schists. These porphyroblasts cut the schistosity of the pelitic schist west of Jerome Brook and in the western part of Corney Brook (Fig. 11), but elsewhere they have been augened and rotated into the schistosity (Fig. 12). At staurolite grade and higher, the rocks tend to develop perceptible compositional banding, and hence to grade toward rocks like the Pleasant Bay complex.

With increasing metamorphic grade, minerals indicative of various grades appear successively, but because of problems with exposure and structural complications, no single section gives a complete set of index minerals. A composite section from Trout Brook and Corney Brook suggests that the order of appearance of porphyroblasts with increasing grade is chlorite, chloritoid, garnet, biotite, staurolite, cordierite, andalusite. The significance of this sequence is discussed in the section on Metamorphism.

Contact relations

Many of the outcrops of pelitic rocks are very rich in chlorite, and in some instances the pelitic rocks are interbedded with mafic chlorite schists. These relations suggest a close relationship between the volcanic and pelitic parts of the Jumping Brook complex. Few direct contacts have been found involving these rocks. At high grade pelitic schists commonly contain lenticles, commonly boudined, of reddish, muscovite-bearing pegmatite, perhaps correlative to unit SDg, but differing from it in their lack of lineation and fine grained character. No dykes have been observed, transposed or not, with the possible exception of the gabbroic material on Faribault Brook. At lower grades the unit commonly contains lenticles of quartz, but they are not useful for correlation. Along the coast of the Gulf of Saint Lawrence from Pillar Rock to Corney Brook rocks of the Jumping Brook complex are intensely cataclasized, and consist essentially of very strongly and thinly cleaved mylonites. The contact between recognizably mylonitic schists and less deformed material lies somewhere east of the Cabot Trail, but is poorly defined.

Age

The age of this unit is not known except insofar as it is associated with the volcanic unit emplaced at 439 ± 7 Ma and metamorphosed at 369 ± 9 Ma (Currie et al., 1982).

Correlation

This unit was correlated with the George River Group by Milligan (1970). The objections to this correlation have already been outlined. However Milligan (1970) felt that the

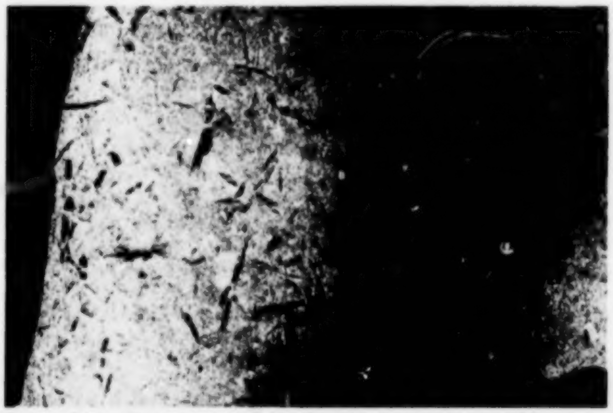


Figure 11. Randomly oriented chloritoid porphyroblasts in pelitic horizon. The outcrop goes around 90 degrees of arc on the left side of the photo, so that the porphyroblasts are unoriented in any of the three dimensions. Trout Brook, 1 km from mouth. (GSC 190886)

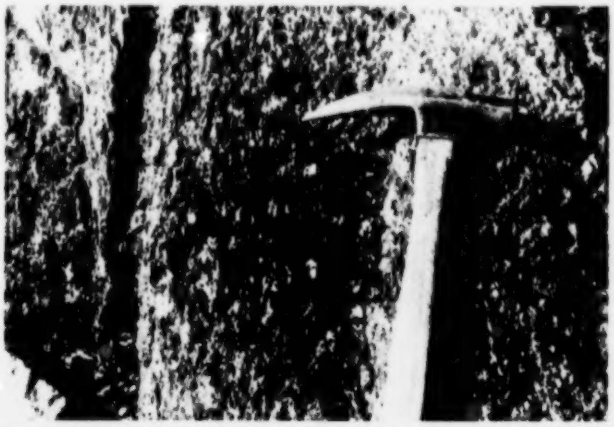


Figure 12. Knotted schist with porphyroblasts of garnet, cordierite and andalusite elongated along the schistosity Corney Brook, 4 km from mouth. (GSC 188588)

unit was correlative to lithologically similar rocks in the Middle River area to the south, a more reasonable correlation which has been supported by Jamieson and Doucet (1983). The pelitic rocks closely resemble descriptions by Wiebe (1972) of pelitic rocks from Clyburn Brook, 15 km east of the mapped area. These rocks also contain garnet, staurolite and andalusite at appropriate grade, and the metamorphic assemblages differ markedly from the surrounding older gneisses. Pelitic schists in the Cape North region may also be correlative, but MacDonald and Smith (1980) suggested that the phyllites of the Money Point Group, which contain cordierite, chloritoid and staurolite at appropriate grade, are correlative to the higher grade gneisses of the Cape North complex which contain kyanite.

Psammitic schist (unit Sjbg)

Definition and distribution

The largest part of the Jumping Brook complex consists of fine grained quartz-rich felspathic rocks which occur as dark grey-green to pale grey, thinly bedded rocks in the lower grade parts of the complex, and as more massive grey to white, 'dirty' quartzites in the high grade parts of the complex. From the upper part of Jumping Brook south to Chéticamp River greenish to greyish quartz-rich metasedimentary rocks form about 70 per cent of the central part of the Jumping Brook complex. South of Chéticamp River similar quartz-rich rocks occur interbedded in the pelitic schists, but they are less dominant.

Lithology

In outcrop the psammitic rocks tend to be finely layered, commonly with bed thickness of less than 1 cm. Foliation tends to be very even and regular, and commonly appears due to alignment of metamorphic minerals along cleavage, but locally on Corney and Robert brooks there are suggestions of crossbedding and grading, so that the foliation may locally be of composite origin.

The rocks are fine grained and homogeneous. They contain sufficient muscovite and/or chlorite to give the rocks a phyllitic sheen. Unlike the pelitic schists, porphyroblasts are rarely conspicuous, although garnet and biotite rarely appear as euhedral porphyroblasts. A few outcrops on the Cabot Trail south of Corney Brook exhibit moderately elongated fragmants of quartz up to 2 cm across. These localities may represent deformed conglomerates. They lie along the edge of a cataclastic zone, in which the Jumping Brook complex has been reduced to mylonite, and confident correlation of horizons is prevented.

Contact relations

Few contacts were seen between the psammitic rocks and other units. The psammitic rocks are interbedded with pelitic horizons of unit Sjbp along Faribault Brook and in the upper parts of Corney Brook and Robert Brook. Amphibolitic dykes assigned to unit Sjbv locally cut quartzite in Faribault Brook. In the upper parts of Corney Brook, Robert Brook, and at the mouth of Mackenzie River quartz-rich biotite-muscovite psammites of this unit appear to grade into a fine grained, white to pale grey, gneissic, biotite-muscovite granite (unit SDg). This granite locally veins the psammitic matrix, and contains within it blocks and enclaves of the schist and gneiss. In general the transition is so gradual that delineation of boundaries between the psammite and the leucocratic granite is quite arbitrary. Field evidence suggests that this granite is some type of metamorphic equivalent of the psammite, perhaps derived by partial melting. Along the lower course of Mackenzie River a mass of this granite forms the transition from the Jumping Brook complex to the Pleasant Bay complex, and seems to grade towards both metamorphic units.

Age

The age of the psammite is not independently known, but presumably must be similar to the U-Pb age on zircon obtained for the volcanic unit of the Jumping Brook complex.

Correlation

Psammitic rocks have been reported from all of the possible correlatives of the Jumping Brook complex (Jamieson and Doucet, 1983; Wiebe, 1972; MacDonald and Smith, 1980), although they are described as relatively minor in the Clyburn Brook sequence (Wiebe, 1972). The same difficulties concerning basement-cover relations apply to correlation of the psammites as bedevil correlation of the other parts of the Jumping Brook complex. In the author's opinion all of the above mentioned sequences are correlative, are of Paleozoic age, and were metamorphosed in the Devonian, as indicated by radiometric age dating of the Jumping Brook complex. A different opinion, however, has been expressed by Jamieson and Craw (1983) and MacDonald and Smith (1980), namely that these metasedimentary sequences are Precambrian and were metamorphosed in late Precambrian time. A necessary corollary of this view is that Devonian metamorphism was much less intense than late Precambrian metamorphism. To the extent that the psammitic rocks grade into, and are migmatized by granites of Devonian age (unit SDg), the psammites of the Jumping Brook complex indicate that the concept of weak Devonian metamorphism is incorrect.

Foliated white, biotite-muscovite granite (unit SDg)

Definition and distribution

Pale buff to white, homogeneous but foliated, fine grained leucocratic granitoid rocks occur commonly in the upper parts of Corney Brook, Robert Brook, and near the mouth of Mackenzie River. These rocks are invariably associated with quartz-rich psammitic rocks of unit Sjbg, and in many outcrops the boundary between them can be detected only with difficulty. The unit is defined therefore as fine grained, homogeneous but foliated granitic rocks. This unit consistently contains muscovite in contrast to the granitic gneisses of unit Hpbg and can therefore be confused only with the muscovite leucogranites of unit Dl. The latter unit however is typically red and coarse grained to pegmatitic, so that confusion does not arise in practice.

Lithology

In outcrop the foliated biotite-muscovite granite is homogeneous and fine grained, commonly with an aplitic texture giving a superficial resemblance to feldspathic quartzite. The rock commonly fractures into angular blocks giving outcrops a rubbly appearance. On weathered surface they may appear almost massive, but on fresh surfaces a pervasive foliation due to alignment of micas, and some elongation of quartzo-feldspathic constituents is visible. In some outcrops a strong alignment of biotite spindles gives a distinct L > S fabric. In the upper part of Robert Brook the granitoid rocks form lit-par-lit gneiss with pelitic schists in layers of 20-100 cm thickness.

Along the Cabot Trail at the bridge over Mackenzie River, the granite forms a rather nebulous agmatitic gneiss with partially assimilated blocks of biotite-rich material forming elongate lenticles along the gneissosity. The granite along Mackenzie River contains fluorite as disseminated grains and small patches.

Contact relations

The foliated white biotite-muscovite granite displays few crosscutting relations. Occasional lenticular masses appear to cross the gneissosity of psammites and pelites of the Jumping Brook complex, but these relations are very local, and generally ambiguous. Gradational relations with the psammites are common, and the gradation appears to occur over various widths from ten to several hundred metres. A gradation to hybrid gneiss occurs east of Mackenzie River, where the foliated biotite-muscovite granite gradually acquires gneissosity, and redder, coarser grained biotite granitoid rocks appear in the gneissic bands. My mapping puts the boundary of the homogeneous granite considerably farther west than the work of Neale (1963), but the exact position of the contact requires some arbitrary decision. Fragments of the foliated biotite-muscovite granite occur in abundance in conglomerates of the Horton Group along the Cabot Trail south of Mackenzie River.

Age

Cormier (1972) obtained a Rb-Sr isochron age of 364 ± 7 Ma on a sample of foliated muscovite granite transitional to composite agmatitic gneiss from the Cabot Trail south of Mackenzie River. This age, based on only two points, is essentially indistinguishable from ages of 367 Ma and 413 Ma from biotite and muscovite respectively obtained by Fairbairn et al. (1960) on a specimen from the same general locality. These ages strongly suggest that the granite is of Devonian age, and hence suggest intensive metamorphism and anatexis in the Devonian.

Correlation

Few rocks possibly correlative to the foliated biotite-muscovite granite have been described from the Cape Breton Highlands, partly because few studies of the granitic rocks have been made. Wiebe (1972) mentioned no muscovite-bearing plutons east of the mapped area, foliated or not, nor did Jamieson and Doucet (1983) mention muscovite-bearing rocks to the south of the mapped area. MacDonald and Smith (1980), however, reported rocks very similar to those considered here in the Cape North region, to the northeast of the mapped area. Their Gulch Brook pluton is foliated, contains minor muscovite and myrmekite, and is thought to be related to the formation of an "injection complex". These characteristics closely resemble those of the foliated biotite-muscovite granite, but MacDonald and Smith (1980) considered the Gulch Brook pluton to be Precambrian. They, however, were unable to distinguish geochemically between this pluton and plutons of Devonian age. Age dating results on the Gulch Brook pluton yielded an essentially two-point isochron age of about 458 Ma by Rb-Sr (Cormier, 1980) with mica ages of 347-361 Ma. These mica ages are essentially identical to those obtained for the foliated muscovite-biotite granite at Mackenzie River, but the isochron age is similar to that assigned to the Jumping Brook complex. These results would be compatible with anatexis of Ordovician to Silurian rocks during the Devonian, as hypothesized for the Jumping Brook complex.

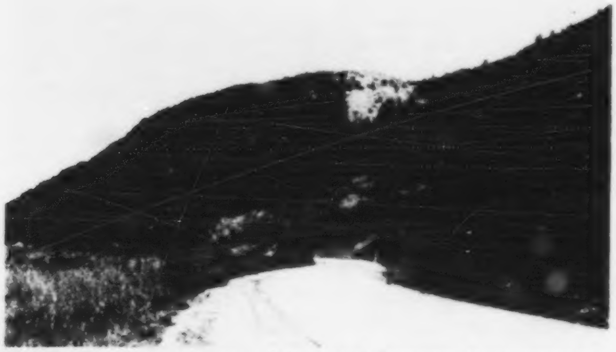


Figure 13. Large mass of lineated leucocratic muscovite-garnet pegmatite and aplite cutting hybrid gneiss. Cabot Trail north of Corney Brook. (GSC 188570)

Leucocratic muscovite granite (unit Dl)

Definition and distribution

A distinctive coarse to pegmatitic, pink to red muscovite granite outcrops as dykes and lenses within the composite granitic gneiss (unit Hpbc). These rocks are well exposed in roadcuts on the Cabot Trail (Fig. 13 and 14) from Corney Brook up the valley of Jumping Brook, and at the head of Faribault Brook. Similar, but poorly exposed examples outcrop along the Cabot Trail from Fishing Cove River northward, in Mackenzie River and in the highlands to the north of Mackenzie River. This unit seems to be as characteristic of the hybrid gneiss as the foliated biotite-muscovite granite (unit SDg) is of the psammites of the Jumping Brook complex. Like that unit, there is a tendency for the bodies to occur near the contact between the schistose and gneissic units. Many of the bodies are too small to show in Figure 2, but it is estimated that the composite gneiss contains about 5-10 per cent of this unit by volume. The lineated muscovite leucogranite forms a very distinctive and easily recognized unit characterized by coarse to pegmatitic grain size, strong pink to red colour, and absence of dark minerals.

Lithology

In outcrop the rocks have a massive, blocky appearance (Fig. 13). In many cases the leucogranite clearly crosscuts the gneissosity of the surrounding rocks (Fig. 14). Although superficially massive, the rocks all show strong lineation, and in many cases foliation, parallel to the local trend. The structural elements are defined by alignment of coarse aggregates of feldspar and quartz, and by lenticular grains of quartz (leaf texture). In lineated rocks these aggregates form spindles up to a centimetre across and several

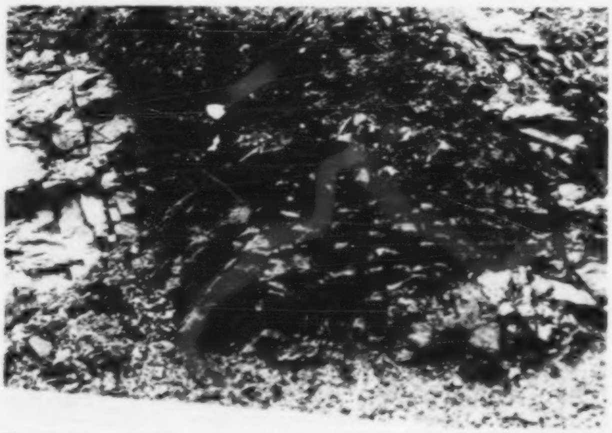


Figure 14. Dyke of red leucocratic muscovite-garnet pegmatite cutting schists of the Jumping Brook complex. Cabot Trail along Jumping Brook. (GSC 188574)

centimetres long, whereas in the foliated rocks the aggregates have a platy character. In both cases sections perpendicular to the lineation may show graphic texture. Mafic minerals are rare or absent, except for reddish garnet which commonly occurs within muscovite books as euhedral crystals. Large books of muscovite occur in the partings between salic aggregates. Some bodies exhibit macroscopically visible mortar texture defined by a tracery of white lines around large feldspar grains. Despite rather heterogeneous texture, varying from moderately coarse grained to pegmatitic within a few tens of centimetres, the rocks appear to be remarkably consistent in composition and character.

Contact relations

The lineated muscovite leucogranites clearly crosscut the gneissosity of the hybrid gneiss (**Hpb**) in most places but in some cases the leucogranites appear to form lenticles within the gneissosity. Where the leucogranite crosscuts the composite gneiss, foliation in the leucogranite is parallel to that in the surrounding gneiss, and hence oblique to the edges of the leucogranite. Contact relations have not been observed with any other unit. No mafic dykes have been observed cutting the leucogranite.

Age

No radiometric age measurements have been made on this unit, and as contact relations are not known, no reliable estimate of the age can be made. The unit, however, shares several characteristics with the foliated biotite-muscovite granite (unit **SDg**) notably presence of muscovite, foliated rather than gneissic character, locally crosscutting relationships, and confinement to a single rock unit. These relationships may be due to similar age. In particular the unit looks only slightly less fresh and massive than the Devonian Salmon Pool and Margaree granites (units **Dsp** and **Dm**). These considerations, as well as the known Devonian age of the foliated biotite-muscovite granite suggest a Devonian age for the leucogranite.

Correlation

To the author's knowledge, no similar rocks have been described from the Cape Breton Highlands, although Milligan (1970, p. 49) mentioned a muscovite pegmatite about 30 km south of the mapped area. Reconnaisance along bush roads suggests that this lack of correlatives results purely from lack of mapping. Similar muscovite leucogranite or pegmatite occurs both north and east of the mapped area in roadcuts, although most of the bodies would be too small to show on 1:50 000 scale maps. The association of muscovite leucogranite with the hornblende-bearing hybrid gneiss probably has petrologic significance, since experimental studies suggest that melts in equilibrium with hornblende-bearing rocks are more potassic than those in equilibrium with similar compositions which are not hornblende-bearing (Helz, 1976).

The rocks resemble some of the muscovite-garnet leucogranites found in the Gander zone of Newfoundland (Strong et al., 1974), possibly due to similarities in mode of origin.

Margaree granite (unit Dm)

Definition and distribution

The Margaree pluton forms a roughly linear mass extending south from the headwaters of the Grande Anse River across the Chéticamp River and south of the mapped area. The continuity of the pluton is interrupted south of the Chéticamp River by faulted zones which bring a rib of composite gneiss (unit **Hpb**) and granite gneiss (unit **Hpbg**) across the trend of the pluton. I have also assigned to the Margaree granite a number of spectacularly porphyritic granitic dykes that cut the composite gneisses around the pluton. Jamieson and Craw (1983) showed some of these dykes as belonging to the Fisset Brook Formation (unit **DMFb**).

Although distinctive and abundant, the Margaree granite is difficult to sample and study in detail because of difficult access and sparse outcrop. Within the map area the rocks can be examined only by difficult climbs to the headwaters of Grande Anse and Mackenzie rivers, or by long walks on the trail from Fishing Cove River to Ingonish. The southernmost exposures are presently easily accessible on a network of woods roads built by Nova Scotia Forest Industries, which are rapidly deteriorating. Outcrop of the granite is virtually non-existent except in stream valleys, but the area of occurrence can readily be mapped from the characteristic feldspar gravel that results from decomposition of the rock.

The Margaree granite consists of euhedral to subhedral alkali feldspar tablets up to 10 cm in length set in a matrix of coarse grained quartz, biotite and feldspar (Fig. 15). This megacrystic texture is highly characteristic.

Lithology

In outcrop the Margaree granite is pink to buff, but weathers to a pale chalky appearance. The distinctive feldspar megacrysts commonly form more than half the volume of the rock, although their abundance locally falls as low as 10-15 per cent, particularly near the margins of the body where the megacrysts tend to be sparse and small. In gross aspect many of the megacrysts look euhedral, but in detail they are rounded, or even blobby. Their size varies markedly both within and between specimens, but there is everywhere a marked distinction in grain size between matrix and megacrysts. In some outcrops the megacrysts are a brilliant red and look distinctly altered. Rims of plagioclase (rapakivi texture) are megascopically visible on many megacrysts. Megacrysts commonly contain many small, dark inclusions which may be zonally arranged.

The matrix of the rock consists of coarse grained bluish quartz, lustrous black biotite, and a pale buff feldspar. Traces of purplish fluorite are a common accessory, and

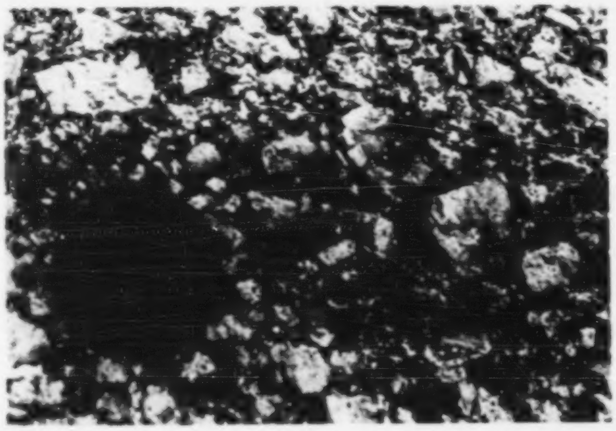


Figure 15. Megacrystic texture in the Margaree pluton. Note thin white rims (plagioclase) on some megacrysts. Headwaters of North Aspy River. (GSC 188590)

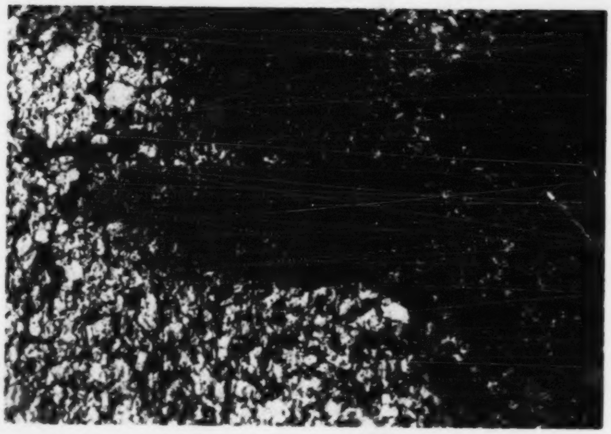


Figure 16. Nebulous amphibolitic inclusion in Margaree pluton. Note megacrysts within the inclusion.' North Aspy River. (GSC 188591)

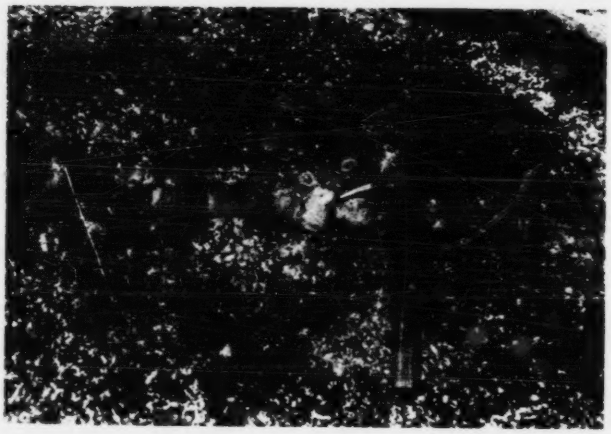


Figure 17. Dyke of biotite granite (Salmon Pool granite) cutting megacrystic Margaree pluton. Canyon of Chéticamp River. (GSC 188592)

small amounts of hornblende occur in the southern part of the pluton. Border phases of the pluton are much finer grained, and approach quartz-feldspar porphyry. This transition can be examined south of Mackenzie River. Within 100 m of the contact with the hybrid gneisses the grain size of both megacrysts and matrix decreases, and within 10 m of the contact the rock becomes a fine grained red porphyry. Dykes of similar material occur in the surrounding gneisses.

The Margaree granite is essentially massive. In a few outcrops a faint alignment of feldspar tablets can be measured, but this foliation cannot be traced from one outcrop to another. Near the margin of the mass, particularly on the headwaters of North Aspy and Northeast Margaree rivers, the rock is crushed, altered and granulated, presumably due to faulting.

Contact relations

The Margaree granite intrudes composite gneisses of the Pleasant Bay complex (unit Hpb). The granite is reddened and chilled against its host, whereas the host gneisses are reddened and locally remobilized with crumpling and loss of gneissosity. Assimilated inclusions of the gneiss occur in the margins of granite (Fig. 16). In the upper part of Grande Anse River, syenites of unit Hm appear to be reddened and fractured against the Margaree granite, but no distinctive contact relations were seen. In the upper part of Red River, a dyke which I attribute to the Margaree granite, but which Neale (1963) correlated with the Fisset Brook Formation, cuts the anorthosite. No relations between the Margaree granite and the Jumping Brook complex were found in the map area, but Jamieson and Craw (1983) mapped the pluton as cutting off the southerly extension of the Jumping Brook complex. The Margaree granite appears to be at least slightly older than the Salmon Pool granite, since dykes of the latter are found in the former (Fig. 17). Neale (1963) interpreted a dyke of Fisset Brook equivalent to intrude the Margaree granite at the headwaters of Grande Anse River. I believe this dyke is a slightly younger phase of the Margaree granite, but it may be equivalent to the Fisset Brook Formation, as suggested by age dating results. No distinctive boulders of the Margaree granite have been recognized in the Horton conglomerates, but many of the abundant, small red feldspar fragments may come from this source.

Age

Jamieson and Craw (1983) reported a Rb-Sr isochron age of 365 ± 46 Ma for the Margaree granite just south of the map area. This age is essentially identical to those obtained for the muscovite-bearing granites (units **SDg** and Dl) and suggests widespread Devonian plutonic activity. The age refutes the suggestion of Keppie (1979) that this granite is of Proterozoic age.

Correlation

The Margaree granite petrographically closely resembles the Cameron Brook pluton described by Wiebe (1975) some 20 km to the east of the mapped area. That pluton, however, appears to be essentially circular, whereas the Margaree pluton is strongly elongate. The age of the Cameron Brook pluton is not definitely known, but Wiebe (1975) interpreted it to be of Acadian (that is, Devonian) age. As noted above, several dykes correlated by me with the Margaree pluton were assigned by Neale (1963) to what is presently termed the Fisset Brook Formation. This suggests that the Margaree pluton may be correlative with the acid parts of the Fisset Brook Formation, a suggestion which is supported by Rb-Sr isochron dates of 376 ± 12 and 370 ± 20 Ma on the Fisset Brook Formation (Cormier and Kelly, 1964).

Megacrystic granites are a relatively uncommon rock type in the Canadian Appalachians, but occur commonly in the Gander zone of Newfoundland (Jayasinghe and Berger, 1976). In that region megacrystic granites of more than one age occur, as they do in the Chéticamp region, since parts of the older Cheticamp granite are also megacrystic. Megacrystic granites also occur in central New Brunswick (Rast and Lutes, 1979). Presumably these unusual rocks represent a particular environment of emplacement and/or solidification along with metasomatism. However the conditions specified by this texture remain poorly understood. It has been suggested (Neale, 1963; Currie et al., 1979) that such textures require participation of basement in the formation of the mass.

Salmon Pool granite (unit Dsp)

Definition and distribution

Pale pink to red, inequigranular to porphyritic biotite granite forms two major plutons within the map area, as well as numerous small stocks and dykes within the composite gneiss (unit Hpb). Both plutons outcrop in the valley of the Chéticamp River, and the name is chosen from particularly striking outcrop at the first Salmon Pool on the Chéticamp River about 3 km east of the Cabot Trail and, readily accessible by a walking trail in Cape Breton Highlands National Park. The rock-type is defined by its massive, medium grained, relatively potash feldspar-rich character. Biotite is the only mafic mineral identifiable in hand specimen, and commonly forms elongate spindle-like masses which show no alignment.

Lithology

In hand specimen the rocks are rather fine grained and markedly inequigranular to porphyritic. Many have aplitic matrices, with larger crystals of potash feldspar. The pluton exposed on Artemise Brook contains potash feldspar euhedra up to 1 cm in length. Although many of the specimens contain feldspar crystals markedly larger than the surrounding material, true porphyritic texture, that is a markedly bimodal distribution of crystal sizes, is relatively uncommon, since most of the specimens have a seriate distribution of sizes. The rocks generally contain very few mafic minerals, and approach alaskite, although local regions have colour indices as high as 20.

The texture of the Salmon Pool granite is notably massive. Alignment of crystals seems to be totally lacking in this unit, and cataclasis is observed only around the extreme southern exposures on LeBlanc Brook.

Dykes attributed to the Salmon Pool granite occur widely in all older units. Such dykes have the rather variable pink to red shade of the plutons, completely massive textures, and seriate grain sizes, dominated by a relatively fine grained aplitic fraction. Only dykes of this character cut the gneissosity of the composite gneiss of the Pleasant Bay complex. Such dykes can be readily distinguished from those attributed to the Margaree granite by their lack of porphyritic character.

Contact relations

Contacts of the Salmon Pool granite are poorly exposed, and most of the available information comes from dykes. At Salmon Pool itself an igneous breccia contains fragments of the Jumping Brook complex in a matrix of Salmon Pool granite. In the canyon of Artemise Brook, the eastern contact of the pluton is partially exposed, and the granite forms an intrusion breccia with granitic gneiss (unit Hpbg) of the Pleasant Bay complex. At both Salmon Pool and Artemise Brook the fragments in the intrusion breccia show no evidence of hornfelsing, but rather are unusually coarse-grained, and appear partly assimilated and metasomatized.

Dykes of the Salmon Pool granite occur in Pleasant Bay complex, Cheticamp pluton, Jumping Brook complex, and Margaree granite. Dykes cut these units with sharp planar contacts (Fig. 17), with only minor chilling and little visible hornfelsing or alteration of their wall rocks.

Fragments of the Salmon Pool granite occur prominently in the Horton Group clastic rocks (unit MHs) south of the Chéticamp River.

Age

The age of the Salmon Pool granite is not directly known. Crosscutting relations with the Margaree granite, as exposed by dykes, and implied by Figure 2 and magnetic patterns in Artemise Brook, imply that the Salmon Pool granite is younger than the Margaree granite, although the difference in age may be minimal. A petrographically similar, possible correlative "younger granite" in the Cape North region to the northeast (MacDonald and Smith, 1980), gave a Rb-Sr isochron age of 330 ± 23 Ma (Cormier, 1980). If this Carboniferous age is accepted, the Salmon Pool granite would be younger than the Fisset Brook Formation, which is dated by both radiometric and fossil evidence as late Devonian to Mississippian.

Correlation

The Salmon Pool granite may possibly be correlative to the "younger granite" mapped by MacDonald and Smith (1980) in the Cape North area, but such correlation must be made with caution since these authors recognize only two ages of granite in this region. Similarly, correlation with the granite plutons mapped by Wiebe (1972), or those dated by Cormier (1972) cannot be reliably established because of the inadequate nature of the sub-division of the granites in these works. The Chéticamp region demonstrates that Devonian magmatism was a complex and probably long-lived process, beginning with local anatexis during deformation (map units SDg and Dl) and ending with the emplacement of post-tectonic major plutons (Margaree and Salmon Pool granites). Jamieson and Craw (1983) suggested that Devonian magmatism in the western Cape Breton Highlands was relatively sparse. The present work, however, demonstrates an extensive and complex history of Devonian magmatism, whereas the eastern highlands have produced, as yet, little evidence for a complex magmatic episode in the Devonian.

Fisset Brook Formation (unit DMFb)

Definition and distribution

Kelley and MacKasey (1965) introduced the term Fisset Brook Formation for a sequence of volcanic and sedimentary rocks in western Cape Breton Island which underlie clastic sedimentary rocks of the Horton Group and unconformably overlie older rocks. The concept of the Fisset Brook Formation helps solve several problems in local geology, but the definition given by Kelley and MacKasey (1965) is defective in several respects. The name of the topographic feature referred to is Fiset Brook, named after a well-known local family. Kelley and MacKasey (1965) variously gave the spelling as "Fisset" and "Fissett". I adopt the former alternative as closer to the correct spelling, as did Poole et al. (1970) It seems unfortunate that the rules of stratigraphic nomenclature do not permit the correction of an obvious typographical error. A more substantive objection arises from the lack of discrimination between the Fisset Brook Formation and the Jumping Brook complex in the original work of Kelley and MacKasey (1965).

Rocks of the Jumping Brook complex can be distinguished from the Fisset Brook Formation by (a) presence of a penetrative fabric in the Jumping Brook complex, and (b) absence of amygdules and/or vesicles in the Jumping Brook rocks. In the mapped area no primary volcanic features of any kind can be distinguished in the Jumping Brook complex, but Jamieson and Doucet (1983) reported that primary features can be distinguished in probably correlative rocks 25 km southeast of the mapped area.

The Fisset Brook Formation, defined by the absence of penetrative fabric and the presence of primary volcanic features, extends in two narrow, parallel belts extending southward from Au Coin Brook to the south edge of the map area, and in a curved belt extending south from Pillar Rock to Chéticamp River. In addition, undeformed basaltic dykes, in some cases vesicular, found in the valleys of Chéticamp River, Au and Coin, Fiset and Farm brooks have been assigned to the Fisset Brook Formation.

Lithology

The Fisset Brook Formation consists of mafic volcanic flows, acid flows and pyroclastic rocks, and minor intercalated sedimentary material. No complete, unfaulted section of the Fisset Brook Formation can be measured within the map area, and the section appears to vary laterally from place to place. The best sections in the map area occur in Fiset and Au Coin brooks, but they are remarkably different, even though they are separated by less than 300 m. Presumably these differences reflect faulting as well as lateral facies changes. In Fiset Brook monotonously similar volcanic rocks occur in flows 5 to 15 m thick consisting of black to purplish brown, fine grained, vesicular lava. Flows may be separated by 0.5 to 2 m of red siltstone, red to olive sandstone, and rare organic layers. The base of some flows is marked by reddened rubbly zones and the tops by highly vesiculated zones, locally with pipe vesicles. The tops so defined consistently face westward, although dips are steep, and commonly easterly. Mafic sills within the pile can be identified by chilling on both sides. The central parts of such sills commonly exhibit ophitic texture. Au Coin Brook displays a mainly rhyolitic section. The lower part contains dense, purple, flow-banded rhyolite which passes upward into moderately welded tuff or ash-fall deposits containing flame-like structures (fiamme) as well as partially resorbed quartz with square cross-section (high quartz?). The upper part of the acid section consists of very vesicular, weakly welded, pale grey material similar to pumice, in which a 5 m mafic flow is intercalated. Sections of similar acid and mafic volcanic rocks occur in Farm and Factory brooks, although exposure is poor. The eastern section of Fisset Brook Formation exposed in the upper part of Farm Brook resembles the sections to the west, but contains a larger proportion of sedimentary material. The sedimentary rocks associated with the Fisset Brook Formation consist mainly of red homogeneous siltstone and sandstone in the western belt, but olive sandstones with organic remains occur in the upper part of Farm Brook. Kelley and MacKasey (1965) reported plant debris recovered from these beds. In Robert Brook the siltstones were deformed while soft, forming spectacular folded, boudin-like structures. Kelley and MacKasey (1965) mentioned a conglomerate containing rhyolite pebbles in Farm Brook, but re-examination of the locality suggests that it is more likely a fault breccia along the contact between the Fisset Brook Formation and the Cheticamp pluton.

Dykes assigned to the Fisset Brook Formation occur in the valleys of Chéticamp River, Fiset Brook and Farm Brook. The dykes consistently parallel the tectonic trend (north-northeast) but in detail cut the gneissosity of their hosts. Dyke widths range from 10 cm to about 1 m. The smaller dykes are fine grained and may have chilled margins. Four contained amygdules. The larger dykes have fine grained margins but visibly ophitic-textured cores.

Contact relations

The contacts of the Fisset Brook volcanic rocks are poorly exposed and somewhat ambiguous. South of the map area, on a woods road east of Pembroke Lake, roadcuts show granite of the Cheticamp pluton broken up into blocks about 2 m across and invaded by vesicular mafic volcanics. The edges of many of the blocks exhibit partial melting and streaking on a scale of several centimetres. On the Acadian walking trail in Cape Breton Highlands National Park, about 100 m north of the park garage, a basaltic flow of the Fisset Brook Formation appears to surround a boss of highly deformed Cheticamp pluton, from which it is separated by about 5 m of intensely flow-banded rhyolite containing nebulous schliers of the granite. No contact was found between the flow rocks and the Jumping Brook complex, but dykes assigned to the Fisset Brook Formation cut the Jumping Brook complex in the upper part of Faribault Brook. No unfaulted contact of the Fisset Brook Formation against the Horton Group was observed in the mapped area. Kelley and MacKasey (1965) described the Horton Group as either unconformably or gradationally overlying the Fisset Brook. MacDonald and Smith (1981) described an unconformity between the Fisset Brook Formation and the Horton Group at Lowland Cove, 25 km north of the mapped area. Within the mapped area, cobbles of Fisset Brook lithologies are notably absent from the Horton conglomerates.

Age

Kelley and MacKasey (1965) assigned an Early Mississippian age to the Fisset Brook Formation on the basis of spores recovered from sedimentary intercalations in Fiset Brook. D.C. McGregor (quoted by MacDonald and Smith, 1981), regarded spores from Fisset Brook Formation at Lowland Cove as being Late Devonian (possibly early Tournaisian) in age. Cormier and Kelly (1964) obtained two Rb-Sr isochron dates from the Fisset Brook Formation, one of 376 ± 12 Ma from material on Fiset Brook itself, and the other of 370 ± 20 Ma on material from south of the mapped area in the region of Lake Ainslie. The Middle Devonian ages suggested by the radiometric ages are essentially identical to those obtained from the late to post-tectonic granite, while the fossil ages are distinctly younger. Jamieson and Craw (1983) suggested that the Fisset Brook Formation varies slightly in age from place to place.

Correlation

The Fisset Brook Formation is currently considered to comprise a chain of occurrences along the western and southern edges of the Cape Breton Highlands stretching from Lowland Cove, 20 km north of the mapped area, to Lake Ainslie, and thence northeast to the vicinity of Saint Ann's Mountain (Keppie, 1979; Jamieson and Craw, 1983). These occurrences all consist of mafic and acid volcanic rocks with intercalated sedimentary rocks of Middle to Late Devonian and Early Mississippian age. The relation of the Fisset Brook Formation to the Horton Group is less certain. Kelley and MacKasey (1965) suggested that the Fisset Brook Formation was locally interbedded with Horton strata. MacDonald and Smith (1981) mapped an unconformity at Lowland Cove. Jamieson and Craw (1983) went even further and suggested that emplacement of granite plutons separated the Fisset Brook Formation from the Horton Group. In the mapped area the contact between the two is generally a fault, but on the ridge west of the Cabot Trail extending from Cheticamp River to Pillar Rock, rhyolite appears to be interbedded with, and possibly intrusive into, granite wash typical of the Horton Group. However outcrop is sufficiently poor that the rocks could be a series of fault slivers. No granites have been observed to cut the Fisset Brook Formation in the mapped area, and dykes assigned to the Fisset Brook Formation clearly cut the Devonian Margaree granite. However the order of units suggested by Jamieson and Craw (1983) raises a crucial and generally neglected question, namely the proper correlation of the Fisset Brook Formation and plutonic granitic rocks. As noted above, the radiometric ages of the

Fisset Brook Formation suggest that they are correlative to Devonian granite plutons, and dykes traceable to the Margaree granite have the same character as material mapped elsewhere as Fisset Brook Formation. If Fisset Brook Formation acid volcanics are extrusive equivalents of the Margaree granite, then they are late to post-tectonic igneous rocks of the Acadian orogeny, rather than the opening symptom of a later event, and the Acadian orogeny should be considered to extend further in time than admitted by classical compilations such as Poole et al. (1970).

Poole et al. (1970) considered the Fisset Brook Formation part of the Horton Group, and correlated it with volcanics occurring within the Horton Group in southern New Brunswick. However for the reasons outlined above the status of the Fisset Brook Formation as a unit of the Horton Group seems quite doubtful, and correlations based on this status remain tenuous.

Horton Group (unit MHs)

Definition and distribution

Poole et al. (1970, p. 287) defined the Horton Group as "...a sequence of mainly Tournaisian (early Lower Carboniferous) continental sedimentary and volcanic rocks that rest unconformably on rocks as old as Precambrian and are overlain conformably or disconformably, and rarely unconformably by the partly marine strata of the Windsor Group." Excluding the volcanic rocks for reasons cited above, such rocks outcrop in a more or less continuous belt along western edge of the highlands from the southern edge of the map area north to Pillar Rock. Farther to the north they appear as a series of fault bounded slivers, and in shoals offshore as far north as Pleasant Bay, where they are exposed in the valley of Grande Anse River.

Lithology

Horton Group rocks in this area fall into two well defined facies. The older consists of conglomeratic, or coarse arkosic rocks derived locally. The younger comprises a much more homogeneous medium-grained feldspathic sandstone with coaly partings. The conglomerate facies contains readily recognizable locally derived fragments. In the northwestern corner of the area fragments of quartz-feldspar porphyry similar to the marginal dykes of the Margaree granite, and also to the Fisset Brook Formation abound. At the mouth of Mackenzie River the conglomerate consists entirely of the underlying pale grey muscovite granite (unit SDg) whereas east of Presqu'ile a conglomerate outlier consists of slabs of the Jumping Brook complex, and 300 m to the south in Jerome Brook the rock is composed of feldspathic fragments of the Cheticamp pluton. These conglomerates contain angular to subrounded fragments from granule size to 30 cm in largest dimension. The matrix, generally 10 to 30 per cent of the volume, may contain either carbonate or quartz cement.

The relations between this lower conglomerate phase and the upper sandstone phase can be observed to be gradational in the lower parts of Farm and Factory brooks. However the sandstone phase is best exposed on Presqu'ile, where the rock is pale green to grey with local reddish silty beds and black organic-rich beds. Unlike the essentially massive conglomerates, the sandstone shows excellent flaggy bedding, commonly graded. The sandstone contains recognizable rounded quartz, potash feldspar, and plagioclase grains as well as detrital muscovite, biotite and tourmaline. The material is well rounded and well sorted (grain size 0.3-0.7 mm) except for rare larger grains of potash feldspar. The fabric is grain supported with a sparse carbonate matrix.

Age

The age of the Horton Group has not been determined directly in the map area, except that it is younger than the Fisset Brook Formation and the massive granitoid rocks. Cameron (1948) mapped the material east of Chéticamp as of lowermost Carboniferous age on the basis of fossil evidence from the contiguous map area to the south. Fairly abundant, although badly deformed, plant fossils occur on Presqu'ile, but they were not studied during the present investigation. Paleoniseid fish remains have also been reported from this site (MacLaren, 1955). The age of the Horton Group has traditionally been taken to be Tournaisian to Early Viséan (Poole et al., 1970).

Correlation

Poole et al. (1970) discussed the correlation of the Horton Group in the Atlantic Provinces in some detail. The group has locally been divided into formations. The present subdivisions correspond to units 9a and 9b of Neale (1963), but the upper unit of arkosic red sandstone which he labelled 9c has not been recognized in the mapped area. The strata may be equivalent to the Craignish Formation which forms the lower part of the Horton Group in the Mabou-Antigonish basin to the southwest (Poole et al., 1970, p. 289), but this remains uncertain. Poole et al. (1970) emphasized that detailed correlations have rarely worked well over long distances in the Horton Group and correlative strata.

Windsor Group (unit MWg)

Definition and distribution

Poole et al. (1970) defined the Windsor Group as "...partly or entirely composed of marine strata that conformably to unconformably overlie the Horton Group and overlap pre-Carboniferous rocks." In the mapped area rocks composed of gypsum, anhydrite and associated limestone and red siltstone have been mapped as Windsor Group on the basis of lithologic similarity to the Windsor Group elsewhere (Cameron, 1948; MacLaren, 1955), although no fossil localities are known, and the relations to other units are obscured by deformation. Such rocks outcrop in a narrow belt extending from the Chéticamp River south to the edge of the mapped area and in a small area around Pleasant Bay. Outcrop is generally poor, except in abandoned gypsum quarries and along a few brooks, but the ground underlain by Windsor Group can generally be recognized by the presence of sink-holes developed in the gypsum. Detailed work on the Windsor Group was not attempted during the present study, except for structural observations.

Lithology

The Windsor Group in the mapped area consists mainly of gypsum and anhydrite, with a prominent 10 m band of limestone (Windsor A horizon?) occurring near the contact with the Horton. Thin smears of poorly indurated red siltstone occur locally near this contact. The outcrops of Windsor Group all appear strongly brecciated and deformed, and thus no meaningful sections can be measured. Both the top and the bottom of the section appear to be faulted, and the thickness of the gypsum-anhydrite horizon varies rapidly along strike. The only reliable stratigraphic marker appears to be the much contorted, and brecciated limestone horizon noted above, assigned by Giles (1978) to the Windsor A subzone. Logs from old water wells suggest that this zone is much less contorted to the west around the town of Chéticamp, although this remains uncertain in the absence of outcrop.

Contact relations

Cameron (1948) stated on the basis of evidence gathered elsewhere that the Windsor Group locally disconformably overlies the Horton. The only exposed contacts within the mapped area involving the Windsor Group are faulted contacts against the Horton Group. Just northeast of Grand Lac a hillock of Horton Group rests on Windsor Group along a low angle fault. In Farm Brook inverted Horton Group sandstone rests on intensely flow-folded carbonate of the Windsor Group, presumably also on a thrust fault. At the northern tip of Grand Lac the Windsor Group contacts pale grey sandstone of the Canso Group along a heavily sheared zone, presumably a fault. Cameron (1948) stated that elswhere the Canso Group disconformably overlay the Windsor, but outcrop of this contact is virtually nonexistent in the Cheticamp region. Windsor Group fragments appear locally in the cataclasites of unit Mbr, and gypsum and carbonate of undetermined origin are abundant in this unit.

The high degree of deformation by folding and flow in the Windsor Group contrasts sharply with the generally weakly folded, or monoclinal, sandstones of the stratigraphically bounding Horton and Canso groups.

Age and correlation

No evidence of the age of the Windsor Group was obtained during the present work, but the group is well established to be of Viséan to Namurian age on the basis of fossil evidence (Kelley in Poole et al., 1970). Correlatives of the Windsor Group in Cape Breton Island and mainland Nova Scotia have recently been discussed by Giles (1978), and in a more general context by Poole et al. (1970).

Breccia, mylonite, flaser gneiss (unit Mbr)

Definition and distribution

Cataclastic rocks of diverse age and character occur widely throughout the mapped area, but the most spectacular and best exposed examples outcrop along the Cabot Trail

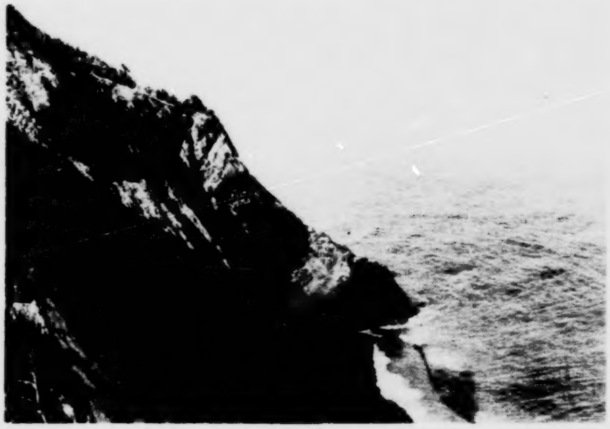


Figure 18. Schist and mylonite developed in the Jumping Brook complex. The foliation visible in the photo is entirely of tectonic origin. An older, probably tectonic, foliation can be seen dipping into the cliff towards the top of the cliff. View south from Cap Rouge. (GSC 188567)

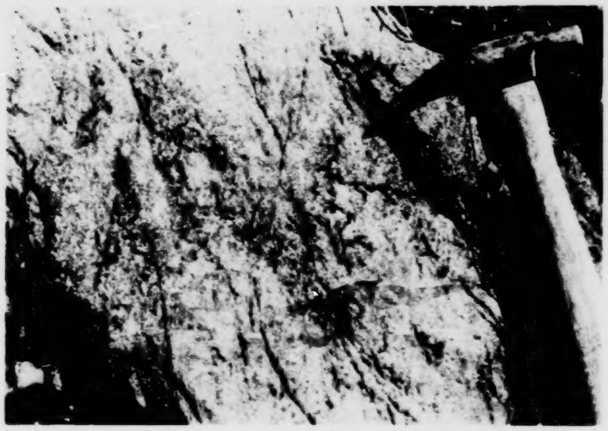


Figure 20. Pseudotachylite veinlet (darker grey under hammer point) cutting breccia and mylonite. Mouth of Corney Brook. (GSC 188571)

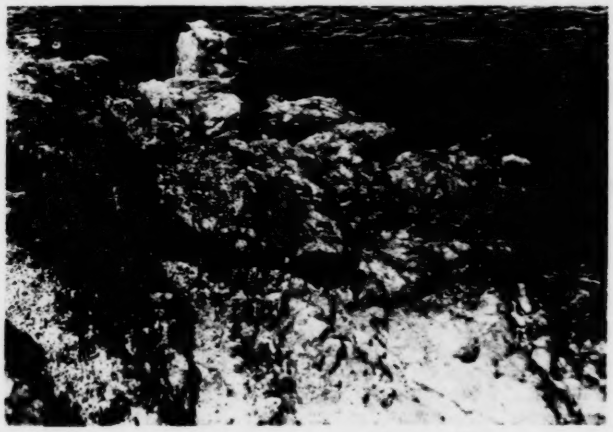


Figure 19. Intensely brecciated Cheticamp pluton, Mouth of Corney Brook. Note the many large fragments, sitting up as lumps, in the intensely ground matrix. (GSC 188568)

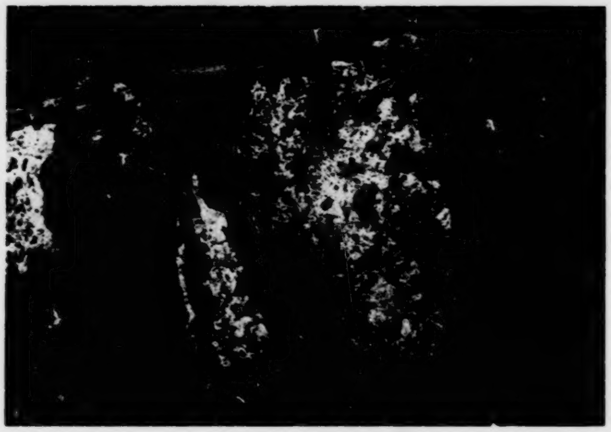


Figure 21. Polymict breccia of fragments of Fisset Brook Formation, Horton Group, and Cheticamp pluton in a quartz-calcite matrix. Robert Brook, 1 km from mouth. (GSC 188597)

from Chéticamp River north to Jumping Brook. Where the protolith of the rocks can be recognized, Figure 2 shows the cataclasis by an overprint, so that they may be assigned to their proper unit. However, many of these rocks have been sufficiently converted by cataclastic deformation that the original protolith can no longer be recognized, and they must be mapped as a separate unit. I have included with this unit all cataclastic rocks sufficiently altered in appearance that they clearly differ from the non-cataclastic parts of the same unit. Such rocks occur in a relatively narrow belt following the sea coast from Pillar Rock north to Jumping Brook, and passing inland along the brook. South of Pillar Rock the zone of cataclasites passes inland, and becomes wider and more diverse. The major zone of cataclasis appears to end at or near the Chéticamp River. Although small patches of similar rocks occur south of the river, no continuous belt was recognized.

Lithology

Four main varieties of cataclastic rocks occur, namely mylonites, homolithic breccias, breccias with pseudotachylite, and heterolithic breccias. Mylonitic rocks appear to have formed mainly from previously banded or stratified rocks, notably the Jumping Brook complex, and seem to maintain in a crude way the previous compositional stratification. Such rocks outcrop from Pillar Rock north to Cap Rouge and again in Jumping Brook. They consist of strongly laminated, pale grey to green, fine grained material with a very strong fissility resembling slaty cleavage (Fig. 18). The surface commonly has a phyllitic sheen, and steeply plunging rodding may be present. In detail, the specimens show flaser structure with extreme elongation and granulation of the small augen. In some cases these augen are themselves fragments of breccias, showing that at least two periods of brecciation were involved. Although the fissility is completely parallel at any one point, over distances of as little as 10 m the attitudes swing by 20 degrees or more. Some outcrops show a strong compositional banding on a scale of a few millimetres, others are essentially homogeneous. None show any sign of porphyroblasts or any relict plutonic textures. Most of the examples appear to have developed from the Jumping Brook complex.

Where the cataclastic zone crosses more massive rocks of the Cheticamp pluton homolithic breccias occur consisting of angular fragments of felsic material from about a centimetre up to several tens of centimetres across. Although a careful examination of fragments shows them to be identical to nearby outcrops of the Cheticamp pluton, on first appearance they seem much more mafic because of the intense chloritization, and concentration of chlorite on fracture surfaces. When struck, this breccia tends to disintegrate into even smaller, irregular jagged pieces, all coated by chlorite. Locally the breccia is strongly reddened due to coatings of hematite. This zone of breccia appears to be an extension of the mylonite zone across massive rocks of the Cheticamp pluton. The breccia has a characteristic lumpy, very crudely banded appearance (Fig. 19) which can be easily identified at a distance.

In the immediate vicinity of Corney Brook brecciation is especially strong, and the rock is reduced to about the texture of a coarse sandstone. This material contains irregular, flame-like dykes and pods of aphanitic grey to black material containing minute feldspar clasts. The general appearance of this material resembles rhyolite (Fig. 20). These flame-like bodies have the appearance and physical character of pseudotachylite. Some possibly analogous material appears along the road to the garbage dump east of Pillar Rock. These exposures strongly resemble flow-banded rhyolite, but may instead be ultra-mylonitic. These very fine grained, flinty rocks constitute only a small part of the cataclastic zone, and appear to have formed along the boundaries of the Cheticamp pluton.

A very different type of breccia appears near Pillar Rock and along Robert, Farm and Factory brooks. The breccia is polymict, and contains abundant matrix, mainly quartz and/or carbonate (Fig. 21). Although the fragments are commonly strongly altered to chlorite, sericite and epidote, their provenance can generally be recognized. Most fragments come from the Cheticamp pluton, the Jumping Brook complex, the Fisset Brook Formation and the Horton Group, with rare fragments from the Windsor Group and the Pleasant Bay complex. The presence of substantial amounts of gypsum in the matrix suggest that it may have been derived from the Windsor Group. Barite is a common, but minor constituent of the matrix.

In addition to these four, reasonably widespread types of cataclastic rocks, there are several interesting though localized examples around Pillar Rock. The rock itself consists of a coarse fault breccia involving mafic volcanics probably from the Jumping Brook complex and a red siltstone of the Horton or Windsor Group. Just to the east, along a high angle reverse fault, occurs an unusual coarse breccia with open voids partly coated by calcite and quartz (Fig. 22) which must be a very late feature. In the hanging wall of this fault, the rocks are veined by gypsum and calcite (Fig. 23) which contain masses of breccia, presumably carried along during emplacement of the vein filling. On Corney Brook, finer grained breccias appear to cut coarser breccias, showing that several episodes of brecciation have occurred.

Contact relations

The cataclastic zones can be seen to cut the Jumping Brook complex and the Cheticamp pluton. In both cases the borders of the zones are gradational, commonly rather abruptly over a metre or so, but occasionally the cataclasis appears and disappears more or less capriciously. Although fragments of the Fisset Brook Formation and Horton groups appear in the breccias, no actual cataclastic zone within these units has been seen. A cataclastic zone probably lies in the lower part of Jerome Brook where roughly flat-lying Horton granite wash to the east abuts against west-facing, overturned, organic-rich Horton sandstone. No cataclastic rocks are known within the Canso or Riversdale groups, which are therefore thought to be younger.

Age

No direct evidence of the age of the cataclastic rocks is known. They clearly must be younger than the Windsor Group rocks, inclusions of which are found within them. However the evidence that the cataclasis is older than the Canso Group rests solely on negative evidence, namely that no Canso fragments have been found in the cataclastics, and the Canso strata appear to be only homoclinally tilted toward the Gulf of Saint Lawrence. The evidence suggests, but does not prove, that much of the cataclasis took place in late Viséan or early Namurian time. Older cataclastic rocks may also occur within the unit, since there is evidence of repeated cataclasis.

Correlation

Tectonic disturbance of Carboniferous strata by folding and faulting is well known in the Atlantic provinces (Maritime disturbance of Poole et al., 1970), but the tectonic regime and the chronology of deformation are still debated (Currie and Nance, 1983). In southern New Brunswick and Nova Scotia deformation commenced in the Viséan and locally outlasted Wesphalian B. In the Chéticamp region cataclasis appears to have been active over a shorter period, since the Namurian Canso Group was not involved. However in a broad sense the Cheticamp rocks may be correlative to some of the intensely deformed material of the Bay of Fundy coast (Currie and Nance, 1983). The author is aware of no description of similar rocks from other parts of Cape Breton Island, with the exception of the mylonite zones described by Jamieson and Doucet (1983) some 25 km south of the mapped area. These zones are said to be cut by undeformed dykes, and therefore appear to be older than the zones north of Chéticamp, at least in terms of latest deformation.

Canso and Riversdale groups (units PMCs and PRs)

Definition and distribution

Olive brown sandstone and red siltstone of the Canso and Riversdale groups underlie Chéticamp Island and a narrow strip along the sea shore. These strata can be readily recognized in outcrop because of their undeformed condition and gentle dips, but outcrop is virtually non-existent except along the sea coast. Fossils characteristic of the Riversdale Group occur at La Pointe on the southern tip of Cheticamp Island (Cameron, 1948).

Lithology

The Canso Group (Mabou Formation according to Cameron, 1948) consists of a lower unit of reddish brown shale, grey and red shale and thin olive sandstones, with an upper unit of grey and red sandstones and red shales. Cameron (1948) tentatively estimated the thickness of the unit to be over 800 m on the southern edge of the mapped area. The Riversdale Group (Port Hood Formation according to Cameron, 1948) consists of massive grey sandstone with an olive cast, rich in plant debris and containing a few coaly partings. Lithologically, the Horton Group on Presqu'ile is essentially identical to the Riversdale Group, but MacLaren (1955) reported a fossil locality on Presqu'ile, implying that it is fixed paleontologically.

Contact relations

No contacts were observed between the Canso and Riversdale strata. They appear to be perfectly parallel, and the contact may be gradational. With the exception of the fault contact with the Windsor strata at the north end of the Grand Lac, noted above, no contacts with older strata were observed. However the relative deformation of the younger and older strata clearly suggests that the Canso and Riversdale strata may be faulted against older rocks or overlie them unconformably.

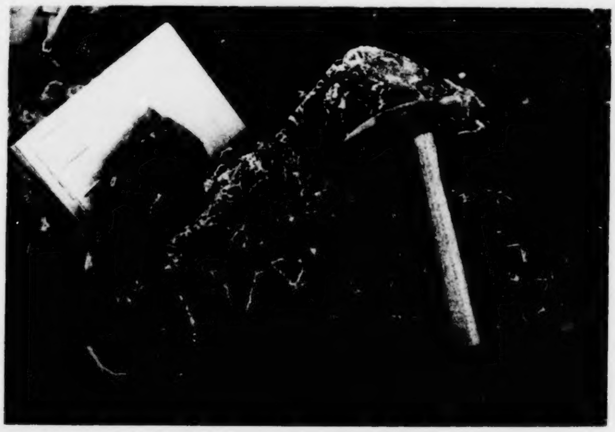


Figure 22. Brecciated Cheticamp pluton with open voids lined by quartz and calcite. Pillar Rock. (GSC 188559)

Age and correlation

Paleontological determinations elsewhere in the Atlantic Provinces show that the Canso Group is of late Viséan to Namurian age, while the Riversdale Group is of late Namurian to Westphalian A age (Poole et al., 1970). The correlation of these groups is discussed briefly by Cameron (1948) for western Cape Breton Island, and by Kelley (1967) and Poole et al. (1970) in more general terms.

METAMORPHISM

Introduction

Pre-Carboniferous rocks of the Chéticamp region have been subjected to metamorphism of diverse types. A working out of the geological history of this region requires an elucidation of the timing and intensity of the various metamorphic episodes. Data gathered in the present study do not permit a complete exposition of the metamorphic history, but clearly show the presence of two quite different and temporally separated episodes of progressive metamorphism.

The periods of metamorphism can be conveniently divided by emplacement of the Cheticamp pluton. Prior to emplacement of this pluton the Pleasant Bay complex had

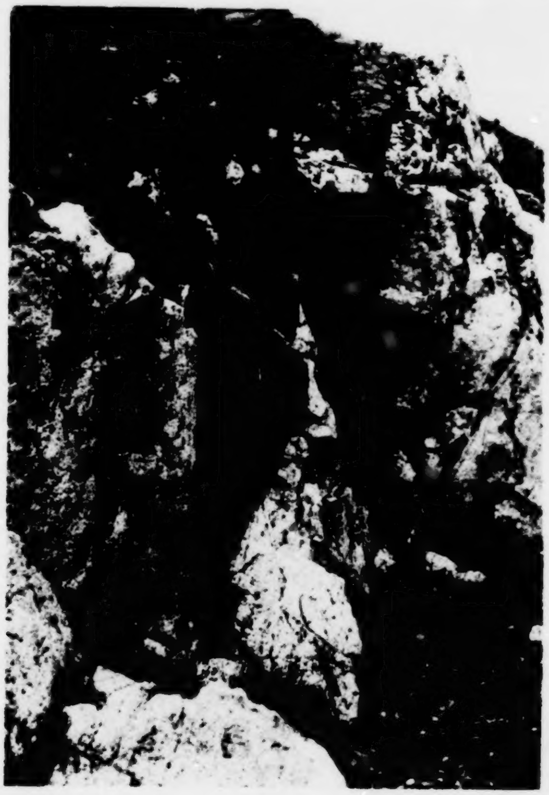


Figure 23. Gash veins of calcite gypsum-barite-fluorite (white) cutting fractured Cheticamp pluton (dark). Hanging wall of reverse fault at Pillar Rock. (GSC 188561)

undergone an extended metamorphic history which can be partly unravelled from a consideration of the mafic dykes and their contained porphyroblasts. Subsequent to emplacement of the Cheticamp pluton and deposition of the Jumping Brook complex the rocks underwent an additional period of progressive metamorphism for which an unusually complete sequence of index minerals is preserved. Finally, all the rocks were subjected to late regressive metamorphism with the formation of chlorite-epidote assemblages.

Metamorphism in the Chéticamp region appears closely linked to magmatic and structural events. This chapter lays the groundwork for a consideration of all these plutonic phenomena together. Petrographic, bulk chemical and mineral chemical data provide a quantitative basis for consideration of the vexed questions of the number of metamorphic culminations and the presence or absence of basement.

Metamorphism of the Pleasant Bay complex

The Pleasant Bay complex consists of three diverse components, namely marble, quartzite and schists and gneisses (Hpbp), amphibolites, demonstrably or presumably of igneous origin (Hpba), and the various granitoid components of units Hpbg and Hpb. Field mapping criteria, summarized in the previous chapter, suggest that the schists and gneisses contain the oldest material. Some of the amphibolitic rocks may represent volcanic material originally associated with these schists and gneisses, but most of the amphibolites appear to have developed from younger intrusive rocks. The various granitoid components of the Pleasant Bay complex appear to be of diverse, commonly relatively young, ages.

Biotite gneiss and schist (unit Hpbp)

The metasedimentary character of parts of this unit can hardly be doubted because of the occurrence of lenticles of marble and quartzite. However much of the schist and gneiss could be developed from either igneous or sedimentary protoliths. The gneiss commonly shows layering on a scale ranging from a few millimetres to 5 cm, although some outcrops show strong foliation without strong compositional layering. The gneisses exhibiting strong compositional layering, have salic bands almost lacking dark minerals. The mafic layers exhibit ubiquitous mineral lineation due to alignment of amphibole, or a crenulation lineation of augen porphyroblasts of feldspar. The gneisses commonly appear coarse grained, and salic lenticles may be pegmatitic, but fine grained, homogeneous foliated rocks occur in narrow arcuate bands less than 30 cm across and a few metres long. These bands may represent recrystallized high strain zones. In addition to the most common homogeneous quartz-biotite-plagioclase (±garnet) gneiss, two other varieties of gneiss form widespread but very minor components of the complex. One is a pale grey to white calc-silicate gneiss readily recognized in the field by the presence of honey yellow garnet. Such gneisses occur as layers up to 10 m wide along the Cabot Trail south of Pleasant Bay, on both sides of the upper Grande Anse River, near the headwaters of Corney Brook, east of the Aspy River near Little Southwest Brook and in the upper part of the Chéticamp River canyon. The other variety of gneiss contains alternating layers of biotite-rich and hornblende-rich material. This composition appears to grade both to the usual biotite gneisses of the unit Hpbp, and to some of the gneissic varieties of amphibolite (unit Hpba). Condensed information on the modes of the gneisses can be found in Appendix 1. All specimens examined microscopically contain quartz, plagioclase and biotite; hornblende, garnet, epidote and sphene may also be present in significant amount, whereas alkali feldspar is rare or absent. In calc-silicate gneiss scapolite, diopside and calcic plagioclase occur.

In the more common gneisses quartz forms polygonal to augen-like grains, commonly clumped together, and shows moderate strain with ubiquitous mosaic texture. Within the fine grained zones quartz dominates the sections, forming a fine grained matrix containing larger cracked and strained grains of quartz. Plagioclase forms rounded to spindle-shaped grains, commonly well twinned, ranging from An_{25} to An_{40} (microprobe and Michel-Levy method). Zoning is negligible or absent. Salic lenticles in the gneiss contain minor tartan-twinned microcline or orthoclase, but no potash feldspar occurs within the mafic-rich parts of the rock. Biotite forms subhedral to euhedral large grains, commonly perfectly aligned, but occasionally showing criss-crossing chains of grains, or defining small intra-folial folds. The biotite exhibits the strong chestnut brown pleochrism typical of biotites of high metamorphic grade. Within the fine grained zones biotite may be heavily chloritized, and some small, ragged grains cross the foliation, as though grown after the strong deformation of this material.

Where hornblende occurs in the gneiss, it forms chains of grains, like biotite, and somewhat resembles biotite in form, but is more poikilitic, containing oikocrysts of quartz and plagioclase. The optic properties (x = yellow, y = olive, x = jade green, $z \wedge c = 27°$) are typical for hornblende, and microprobe analyses (Appendix 3) show that the hornblende contains more than 2 atoms of Al per unit cell, hence differing qualititatively from the actinolitic amphiboles of the Jumping Brook complex. Epidote occurs in almost all specimens of this unit as small rounded granules disseminated more or less randomly. The mineral appears to be in equilibrium with its surroundings, and shows no evidence of reaction. Most of the epidote is pale brown, and shows no anomalous extinction. Some relatively salic biotite gneisses contain no epidote, but instead contain small, dense garnet euhedra. In some cases the garnet forms crescentic masses of small granules, as though an older phase had been broken up and recrystallized (Fig. 24). The mineral shows no trace of the "snowball" habit common in the Jumping Brook complex. Among the accessory minerals, sphene is most common, particularly in amphibole-bearing rocks. It occurs both in the characteristic wedge-shaped forms and as small granules, superficially resembling garnet. Apatite occurs in most thin sections in accessory amounts. Magnetite and ilmenite occur in only two sections. One of the fine grained lenticles contains accessory carbonate. Chlorite occurs in minor amount in several sections, but the textural relations strongly suggest that it is an alteration product of biotite, and not a mineral once in equilibrium with the rest.

The calcareous gneisses occur as tenuous lenticles within the biotite-dominant gneisses just described, but contain a markedly different mineral assemblage. The rocks consist of perfectly aligned lenticles up to a centimetre wide of quartz-rich, diopside + scapolite-rich, and epidote + amphibole + garnet-rich composition (Fig. 25). The quartz-rich layers comprise very strained, coarse mosaic quartz with sutured grain boundaries, possibly relicts of quartz veins. Scapolite-diopside lenticles contain large euhedra of scapolite rich in meionite molecule (Appendix 3) and containing no detectable Cl. Diopside occurs in similar large euhedra and contains about 66 mol per cent of $CaMgSi_2O_6$ end member. The space between these minerals consists of plagioclase varying from An_{65} to An_{80} over short distances, but not systematically, minor quartz, and a little calcite. The epidote + amphibole + garnet layers consist of coarse euhedral epidote with strongly anomalous blue extinction, olive hornblende euhedra, essentially identical to those in the gneisses just described, large amoeboid garnets which are noticeably zoned with Mn-rich cores, and rims rich in grossular component. Seams of biotite zoned from chestnut interiors to pale orange rims occur along the contacts of epidote and diopside bearing lenticles. The texture of these rocks, and the occasional occurrence of all the characteristic minerals together, strongly suggest that equilibrium was attained within and between the lenticles. As in the biotite gneiss, sphene and apatite occur in accessory amounts in the amphibole-bearing part of the rock.

Bulk rock chemical analyses for the modally analyzed specimens of the gneiss are given in Appendix 2. I deliberately avoided obviously metasedimentary rocks such as marble and quartzite, so that the silica-poor compositions implied by these analyses may be somewhat biased. However the bulk-rock composition of the calc-silicate gneiss (specimens 54 and 105) differs surprisingly little from the more common biotite (± hornblende) gneisses. The close association of the former with marble and quartzite strongly suggests that they are of sedimentary origin, but the composition of the biotite gneiss does not strongly suggest sedimentary origin. In order to investigate the origin of these rocks a discriminant function devised by Shaw (1972) for similar rocks from the Grenville province of Ontario was applied to the analyses.

$$\begin{aligned} DF = {} & 10.44 - 0.21SiO_2 - 0.32Fe_2O_3 - 0.98MgO \\ & + 0.55CaO + 1.46Na_2O + 0.54K_2O \end{aligned}$$

where Fe_2O_3 gives the total iron. This function is designed to give positive values for igneous protoliths, and negative values for sedimentary ones. When applied to the analyses of Appendix 2, the function gives negative values for only four analyses (3-1, 21-2, 94, 105), and strongly positive values (greater than 1) for the rest. Shaw (1972) gave the probability of misidentification of a single analysis as about 3 in 10 with this function, although van de Kamp et al. (1976) thought the function less reliable. However using the criteria of van de Kamp (1968) leads to essentially identical results. It therefore appears highly probable that much of the Pleasant Bay complex derives from igneous rocks. Considering the association with clearly sedimentary rocks, it may be assumed that the gneiss (Hpbp) derives from some mixture of sedimentary and volcanogenic material. Shaw (1972) reached the same conclusion for the lithologically similar rock of the Apsley gneiss.

If the igneous provenance of part of the gneiss is accepted, the bulk chemical composition suggests derivation from andesitic to dacitic rocks. The Precambrian age of the gneiss implied by its relation to the Cheticamp pluton precludes derivation from Paleozoic volcanic rocks correlative to those in other parts of the Atlantic provinces, but suggests possible sources in Late Precambrian volcanics such as those of the Avalon zone, or possibly in Hadrynian rocks correlative to the Grenville province to the west, now known to contain substantial amounts of volcanic material (Shaw, 1972; van de Kamp, 1968).

In considering the metamorphic grade of the Pleasant Bay gneisses, the widespread coexistence of oligoclase, hornblende and epidote appears characteristic. According to usual facies series definitions, gneisses containing oligoclase and hornblende lie in amphibolite facies (Turner, 1981, p. 365). Epidote occurs in such rocks only in relatively high pressure parageneses such as those from the Lepontine Alps (Wenk and Keller, 1969). In general terms, the gneisses of the Pleasant Bay complex have been metamorphosed to

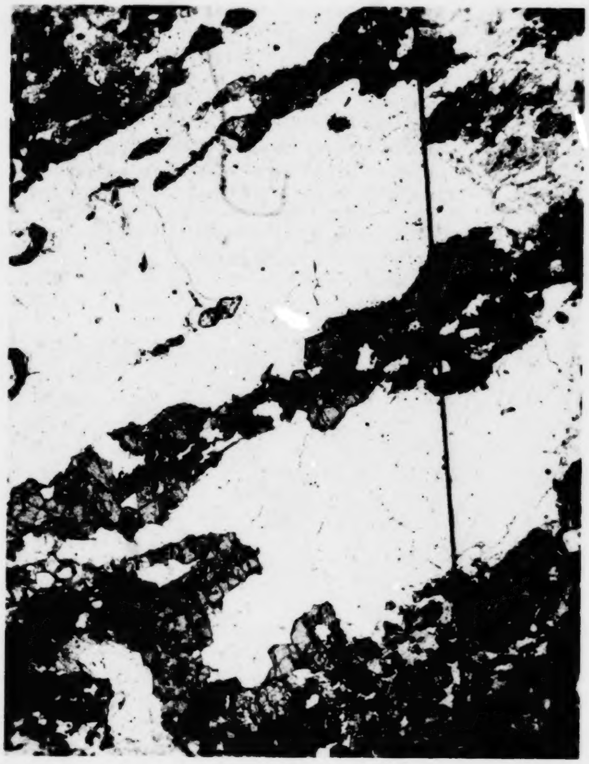


Figure 24. Lenticles of granular garnet (dark grey) between quartz bands and hornblende + epidote-rich bands. Width of field 2 cm. (specimen 54). (GSC 202832-M)

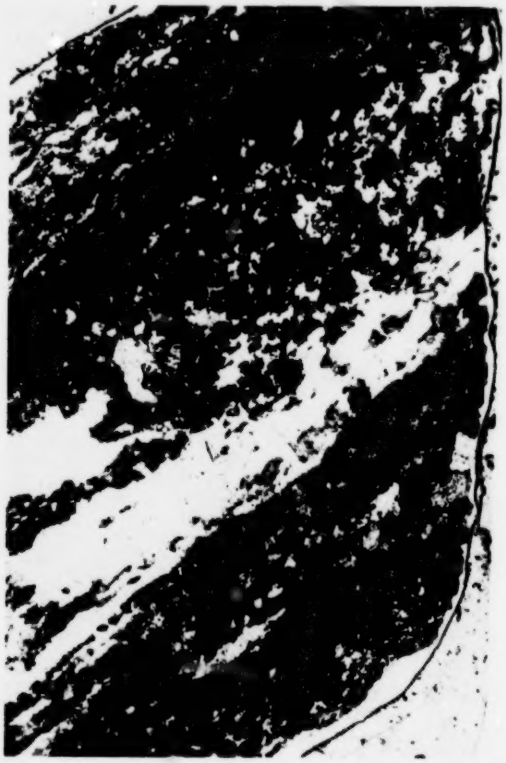


Figure 25. Calcareous granulite gneiss, general view. Note the banded appearance with alternating layers of epidote + hornblende (dark grey) and scapolite (white) + diopside (medium grey). Width of field 3 cm. (specimen 54). (GSC 202832-N)

amphibolite facies at relatively high pressures. The repeated occurrences of the assemblage scapolite-diopside indicates quite high temperature conditions, exceeding 625°C at 500 MPa (5 kilobars), and a mol fraction of CO_2 in the gas phase exceeding 0.12 (Aitken, 1983). The Cl-free, Ca-rich scapolite observed in the calcareous gneisses is particularly characteristic of such conditions. The presence of mafic-poor quartzofeldspathic lenticles at such pressures and temperatures may reasonably be ascribed to partial melting.

The presence of calcareous rocks, the absence of strongly pelitic material, the predominance of relatively mafic gneisses, and the presence of high-pressure, high-temperature assemblages in the Pleasant Bay complex contrast strongly with the pelitic, low-pressure schists of the younger Jumping Brook complex.

Gneissic to schistose amphibolites (unit **Hpba**)

Amphibolites within the Pleasant Bay complex form sharply bounded, homogeneous, but foliated mafic enclaves. In some cases these enclaves form more or less concordant boudin-like masses in the gneiss, but more commonly they are composite bodies consisting of biotite- and quartz-bearing gneiss with smaller, rather nebulous regions of greenish black, homogeneous amphibolite consisting mainly of aligned amphibole prisms. In some cases the amphibolites truncate structures in the surrounding gneisses, and clearly represent transposed dykes (Fig. 5).

In thin section the homogeneous amphibolites consist of chains or networks of aligned amphibole prisms (x = yellow, y = olive, z = jade, $z^{\wedge}c = 20 - 27°$) with the interstices filled by augen-like twinned plagioclase (An_{35-45}) which tends to display weak reverse zoning. Epidote occurs fairly abundantly as rounded brownish granules, in some cases forming a kind of ophitic texture with hornblende. Most specimens contain a little interstitial quartz, and small amounts of aligned biotite occur in strongly foliated rocks. Sphene and apatite are common accessories, but magnetite occurs sparsely. Potash feldspar occurs only as rare porphyroblasts.

The homogeneous amphibolites commonly contain thin lits resembling the biotite gneiss of unit **Hpbp**, or the epidote-rich lits of the calcareous gneiss. Biotitic lits contain augen of microcline (Fig. 26), either of fresh porphyroblastic appearance, or, less commonly, surrounded by salic paste. The biotite-rich seams also contain significant amounts of sulphides, commonly pyrite + pyrrhotite, although small amounts of arsenopyrite occur along upper Mackenzie River.

The gneisses surrounding the homogeneous amphibole-rich rocks consist of mesocratic amphibole-biotite gneiss similar to 21-2 (Appendix 1). These rocks cannot be reliably distinguished from the biotite gneiss (**Hpbp**) to which they grade by loss of amphibole and decrease of colour index. Such rocks quite consistently contain porphyroblastic potash feldspar (Fig. 26), either tartan twinned, or with bluish, shimmering extinction suggesting submicroscopic twinning.

About one third of the rocks mapped as amphibolites contain large amounts of chlorite, particularly along Fishing Cove River. The chlorite appears to be entirely secondary and pseudomorphous after biotite and hornblende. In some rocks chlorite forms a pervasive interstitial phase. North of Grande Anse River chloritization clearly follows late shear planes, but along Fishing Cove River no obvious shearing accompanies chlorite-rich rocks.

Appendix 2 gives seven chemical analyses of amphibolites selected for their fresh, homogeneous character. Rocks displaying obvious potash feldspar porphyroblasts, or numerous intercalations of biotite gneiss were not analyzed, but probably approach the composition of the hornblendic parts of unit **Hpbp** (analyses 45-1 and 90 in Appendix 2). The homogeneous amphibolites display a restricted range of composition, which approaches compositions of average tholeiitic basalt and diorite (Nockolds, 1954), and is quite similar to average diabase from eastern North America (Table 2, columns 3, 4). However the compositions are substantially richer in K_2O than fresh diabase dykes (compare column 1 to columns 3, 4). If such compositions reflect original magmatic compositions, they can be matched only among potassic basalts such as those found on some oceanic islands (cf Pearce and Cann, 1973; Pearce et al., 1975). This seems improbable in the absence of other evidence, and I assume that the amphibolites represent slightly potash-metasomatized diabase dykes. The presence of potash metasomatism is obvious in amphibolites with potash feldspar porphyroblasts, but in the homogeneous amphibolites the "extra" is presumed contained in the amphibole and in the accessory amounts of biotite.

The mineral assemblage of the amphibolites, oligoclase-hornblende-epidote (± quartz ± biotite ± sphene) indicates amphibolite facies metamorphism at relatively high pressure, as pointed out for this assemblage in the biotite gneisses of unit **Hpbp**. Indeed the composition of the minerals in a typical amphibolite (Appendix 3) resembles the composition of the minerals in the surrounding biotite gneiss, suggesting that the two units reached metamorphic equilibrium together. The potash metasomatism responsible for the unusual compositions presumably took place during or before this equilibration, raising the K and Rb content of the rocks, and possibly affecting the Al and Si contents. Since no potash feldspar now occurs in many of the rocks, but removal of a small amount of alkali feldspar composition from the

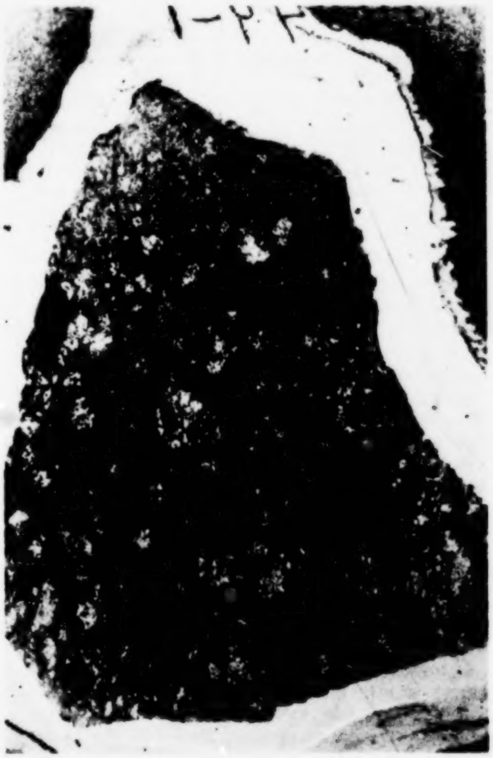


Figure 26. Microcline porphyroblasts in amphibolite. Note that some porphyroblasts are surrounded by haloes of granulated material. Width of field 3 cm. (GSC 202868-A)

Table 2. Chemical affinities of amphibolites of the Pleasant Bay complex

	1	2	3	4	5	6
SiO_2	49.48	48.39	50.83	47.90	51.06	50.40
TiO_2	0.65	0.69	2.03	0.59	0.76	0.44
Al_2O_3	15.53	15.05	14.07	15.26	14.95	15.51
Fe_2O_3	3.51	3.73	2.88	12.10	11.77	0.99
FeO	9.51	10.11	9.00			7.83
MnO	0.28	0.30	0.18	0.18	0.20	0.17
MgO	6.25	6.64	6.34	10.52	7.44	10.60
CaO	9.49	10.09	10.42	10.75	10.80	10.87
Na_2O	2.74	2.80	2.23	2.00	2.23	1.42
K_2O	1.53	0.57	0.82	0.29	0.48	0.37
H_2O	1.27	1.50	0.91	-	-	1.55
P_2O_5	0.13	0.14	0.23	-	-	0.08
Ba	287	-	-	-	-	232
Rb	54	-	-	8	15	230
Sr	151	-	-	115	127	126
Cr	90	-	-	766	218	142
Ni	73	-	-	308	48	85

1. Average Pleasant Bay amphibolite (seven analyses, Appendix 2)
2. Average Pleasant Bay amphibolite less 7% potash feldspar (SiO_2 63.68, Al_2O_3 19.57, Na_2O 1.56, K_2O 14.21)
3. Average tholeiitic basalt and dolerite (Nockolds, 1954)
4. Average eastern North American diabase, ol-normative (Weigand and Raglund, 1970)
5. Average eastern North American diabase, low-Ti, 9-normative (Weigand and Raglund, 1970)
6. Chilled margin Ferrar dolerite, Antartica (Gunn, 1966, p. 890)

rock would return it to a reasonable tholeiitic composition (Table 2), I assume that the introduced alkali feldspar component was incorporated into amphibole and biotite during metamorphism.

Gneissic biotite augen granodiorite (unit Hpbg)

The granitoid gneisses of the Pleasant Bay complex exhibit a characteristic interlocking seriate texture which is not polygonal. This texture appears unique to the Pleasant Bay granite gneisses in this region. The abundant quartz shows moderate lamellar strain. The prominent feldspar augen consist either of microcline with virtually no exsolved plagioclase, or of myrmekite. In either case the composition of the potash feldspar ranges from Or_{85} to Or_{95} with only a trace of Ba (Appendix 3), and no difference between the prominent augen and the granular matrix feldspar. The augen cut cleanly across the foliation of the rock, and show no "wrap around" (Fig. 7), although small amounts of granulated material ("mortar") may be present around the edges of the augen. Independent plagioclase grains occur only in the matrix, where they tend to be markedly more abundant than potash feldspar. The composition of the plagioclase, whether in independent grains or in myrmekite, varies little from An_{20}. In some of the gneisses much of the feldspar is present as myrmekitic intergrowth, and independent grains of either potash feldspar or plagioclase are uncommon. Well aligned biotite, concentrated on certain planes in the rock, tends to be pleochroic in shades of greenish brown, but much of the "biotite" has anomalously weak pleochrism and birefringence, and appears rather heterogeneous and "muddy". Electron probe analysis shows that such material is deficient in K and grades to chlorite. Muscovite occurs in 6 of 8 specimens whose modes were counted, commonly intergrown with biotite but in rarer and smaller flakes. Zircon and apatite occur commonly as accessories. Yellowish epidote and garnet occur in trace amounts in one specimen.

Eight chemical analyses of the foliated granite and granite gneiss of the Pleasant Bay complex have been compiled in Appendix 2. These analyses exhibit a large variation from trondjhemitic (analysis 25) to close to minimum melt compositions. When plotted on normative An-Ab-Or, the analysis give a roughly linear array (Fig. 31, 32) suggesting removal of oligoclase, or addition of a sodic potash feldspar played a major role in the chemical variability. These granitoids resemble the Cheticamp pluton more than other granitoid units because of moderate TiO_2 contents, and significant amounts of Cr and/or Ni even at high silica content. Only one analysis (32-2) contains normative diopside, while 5 contain more than 2 per cent normative corundum, suggesting a distinctly S-type character for the suite as a whole.

Because of the nature of the mineral assemblage of these rocks, and the clearly secondary character of many of the mafic mineral assemblages (chlorite + epidote), quantitative estimates of their pressure and temperature conditions of emplacement cannot easily be made. However the myrmekitic intergrowths which are common in these rocks appear to be ideal for the application of the two

Table 3. Chemical affinities of greenschists of the Jumping Brook complex

	1	σ	2	σ	3	4
SiO_2	49.85	2.37	72.55	3.66	49.48	50.83
TiO_2	1.26	0.37	0.29	0.25	0.65	2.03
Al_2O_3	16.28	0.58	15.10	1.80	15.53	14.07
Fe_2O_3	2.10	1.35	0.92	0.53	3.51	2.88
FeO	6.92	0.52	0.71	1.76	9.51	9.00
MnO	0.20	0.01	0.06	0.05	0.28	0.18
MgO	7.60	1.02	1.58	0.91	6.25	6.34
CaO	9.35	2.22	0.25	1.26	9.49	10.42
Na_2O	3.20	0.88	5.08	0.36	2.74	2.23
K_2O	1.10	0.28	4.19	0.55	1.53	0.82
H_2O	2.48	0.78	0.50	0.49	1.27	0.91
P_2O_5	0.20	0.08	0.10	0.05	0.13	0.23
Ba	172		500		287	
Rb	22		147		54	
Sr	383		72		151	
Cr	168		0		90	
Ni	116		7		73	

1 Average mafic schist of the Jumping Brook complex (6 analyses) and standard deviation (Analyses from Appendix 2)

2 Average acid schist of the Jumping Brook complex (4 analyses) and standard deviation (Analyses from Appendix 2)

3 Average amphibolite of the Pleasant Bay complex (Table 2)

4 Average tholeiitic basalt and dolerite (Nockolds, 1954)

feldspar geothermometer utilizing microcline and plagioclase (Whitney and Stormer, 1977). This geothermometer has the form

$$T(^{o}K) = \frac{(7931.1 - 16910.6x_a + 9901.9x_a^2 + (0.11 - 0.22x_a + 0.11x_a^2)P)}{(1.9872 \ln(x_p/x_a) + 4.6321 - 10.815x_a + 7.7345x_a^2 - 1.5512x_a^3)}$$

where x_a is the mol fraction of albite in alkali feldspar and x_p the mol fraction of albite in plagioclase. If it is assumed that the granite gneiss was emplaced near the peak of metamorphism, the pressures previously deduced would be relevant, that is greater than 500 MPa (5 kilobars). I have assumed a pressure of 700 MPa (7 kilobars). The results of these calculations are shown in Table 4, together with some results for other units of the Pleasant Bay complex. In cases where the results of analyses show zoned or patchy feldspars, I have assumed that the most sodic alkali feldspar originally co-existed with the most calcic plagioclase feldspar. This assumption accords with petrographic evidence. Obvious later overgrowths or patches on the plagioclase are invariably more sodic that the host, and in the case of rims on potash feldspar, they are invariably more potassic. The minimum temperature for a given assemblage can be obtained by using the most sodic plagioclase and the most potassic feldspar. In the case of the Pleasant Bay complex the maxima all fall in the range 640-670°C (913-943°K). This small range strongly suggests that the granite gneiss and the other gneisses were metamorphosed and cooled together, and therefore are correctly mapped as a single metamorphic complex. The temperatures are reasonable for high grade metamorphic rocks, and in accord with those deduced from the assemblage scapolite-diopside. Such temperatures might induce anatexis in favourable compositions, but seem too low for evolution of an igneous granite. The minimum temperatures given by the feldspar thermometer, in the range 300-350°C, presumably reflect late hydrothermal activity forming or altering feldspar.

Metamorphic conditions in the Pleasant Bay complex

The consistent occurrence of the assemblage hornblende-oligoclase-epidote in two units of the Pleasant Bay complex suggests that the observed mineral assemblages result from a single period of metamorphism in amphibolite facies at relatively high pressure, postdated only by late, low-grade retrogression commonly associated with shear zones. Field and petrographic data, however, suggest an earlier metamorphic history now largely obscured by the dominant metamorphic episode. Amphibolitic dykes, now concordant to the foliation, cut migmatitic gneiss which had been deformed before emplacement of the dykes (Fig. 5), but now has mineral assemblages similar to those in the dykes. The deformed and partially remobilized granitoid gneisses are clearly older than the Cheticamp pluton (Fig. 4), which postdates the high-grade metamorphism. The presence of zero, one or two generations of microcline porphyroblasts in amphibolitic layers suggests that more than one period of mafic rock emplacement occurred before the pervasive amphibolite facies overprint. These data point to a relatively complex early history for the Pleasant Bay rocks involving at least two periods of mafic rock emplacement and one period of granitoid rock emplacement prior to emplacement of the Cheticamp pluton.

Some mineral assemblages in the calcareous gneiss suggest granulite facies metamorphism, possibly indicating a link with the granulite facies assemblages noted by MacDonald and Smith (1979) around the Red River anorthosite complex. Mapping and petrographic study demonstrate conclusively that the calcareous gneisses of the Pleasant Bay complex form a single unit with the surrounding hornblende-biotite gneisses, both spatially and metamorphically. There is no justification for separating supposed "George River equivalent" calcareous rocks from the surrounding gneisses, as attempted by Keppie (1979).

Table 4. Temperatures predicted by two feldspar geothermometry

For feldspar pairs it is assumed that the most calcic plagioclase was originally in equilibrium with the least potassic alkali feldspars. For alkali feldspars, temperatures are computed from the formula (Whitney and Stormer, 1977)

$$T°(K) = \frac{(7973.1 - 16910.6\, x_a + 9901.9\, x_a 2 + (0.11 - 0.22\, x_a + 0.11\, x_a 2)}{(1.9872\, \mathrm{lu}(x_p/x_a + 4.6321 - 10.815\, x_a + 7.7345\, x_a\ 2 - 1.5512\, x_a 3)}$$

where x_a is the mol fraction of albite in alkali feldspar and x_p the mol fraction of albite in plagioclase. Derivation of appropriate pressures P is discussed in the Text for monocline alkali feldspar in the formula of Stormer (1972) was used.

$$T°(K) = \frac{(6326.7 - 9963.2\, x \text{ at } 943.4\, x_a 2 + 2690\, x_a 3 + (0.0925 - 0.1458\, x_a + 0.014\, x_a 3)P)}{(1.9872\, \mathrm{lu}(x_p/x_a) + 4.6321 - 10.815\, x_a + 7.7345\, x_a 2 - 1.5512\, x_a 3)}$$

Specimen		x_p	x_a	Order	Pressure bars	Temperature °K	Occurrence
Pleasant Bay complex							
74054	(max.)	0.47	0.12		7000	922	zoned crystals myrmekite
	(max.)	0.70	0.12	o	7000	961	zoned crystals myrmekite
	(min.)	0.92	0.07	o	7000	831	rims on grains
74056-1	(max.)	0.74	0.13	o	7000	958	cores of grains
	(min.)	0.91	0.10	o	7000	874	rims of grains
74107	(max.)	0.77	0.13	o	7000	947	cores of grains
	(min.)	0.94	0.06	o	7000	810	rims of grains
Cheticamp pluton							
74027-3	(max.)	0.65	0.35	d	5000	1128	cores of grains
	(min.)	0.69	0.05	d	5000	689	rims of grains
74061		0.66	0.14	d	5000	661	turbid dusty grains
Jumping Brook complex							
74028-2	(max.)	0.71	0.12	o	4000	921	cores of grains
	(min.)	0.76	0.04	o	4000	767	rims of grains
74008-3		0.76	0.05	o	4000	793	myrmekite
Lineated to foliated granites							
74024-2		0.88	0.07	o	4000	807	graphic intergrowth
74042-2		0.80	0.16	o	4000	927	independent grains
Margaree and Salmon Pool plutons							
74072		0.74	0.10	o	3000	886	cloudy, turbid intergrowth
74100	(max.)	0.65	0.36	d	3000	1058	cores of fresh grains
	(min.)	0.72	0.10	d	3000	892	rims of grains

Table 5. Temperature-pressure estimates for the Cheticamp area

(a) Pleasant Bay complex (specimen 54)

$$\text{garnet: } x_{alm} = 0.42 \quad x_{gro} = 0.39 \quad x_{pyr} = 0.01 \quad x_{spe} = 0.04$$

(alm = almandine; gro = grossular; pyr = pyrope; spe = spessartite)

hornblende Fe/Mg (Hfm) = 0.953 biotite Fe/Mg (Bfm) = 0.844
pyroxene Fe/Mg (Pfm) = 0.493 garnet Fe/Mg (Gfm) = 4.20
plagioclase $x_{an} = 0.78$ scapolite $x_{me} = 0.65$ epidote $x_{cz} = 0.64$

(i) garnet-hornblende geothermometer (Graham and Powell, 1984)

$$T = (2880 + 3280\, x_{gro})/(2.46 + \log (Gfm/Hfm)) = 1064$$

(ii) plagioclase-scapolite (Goldsmith and Newton, 1977)

$$T = (5179 - 9023\, x_{an} + 6518\, x_{an}^2)\; x_{me} - (3695 - 10842\, x_{an} + 8313\, x_{an}^2) = 1073$$

(iii) plagioclase-garnet-pyroxene (Perkins and Newton, 1981)

$$P = 675 + 17.179\; T + 3.5962\; T \log\; (a_{gro}^{2} \cdot a_{pyr}/a_{di} \cdot a_{an})$$

For calculation of activities see text. A temperature of 1069 K is used for calculation. Using the Ganguly parameters for garnet this gives P = 10339 bars. Using the Perkins/Newton parameters for garnet gives 6880 bars. An average value of 8500 bars is assumed.

The following gives results for Fe-Mg exchange geothermometers assuming P = 8500 bars, and using the Perkins/Newton parameters for garnet. The results for Ganguly parameters for garnet are given in parentheses.

(iv) garnet-pyroxene geothermometer (Ellis and Green, 1979)

$$T = (3030 + 0.01086\; P + 3104\, x_{gro})/(1.9034 + \log (Gfm/Pfm)) = 1071$$

(v) garnet-pyroxene geothermometer (Ganguly, 1979)

$$T = (4801 + 0.01107\; P + 1661\, x_{gro})/(2.93 + \log (Gfm/Hfm) + 0.755\, x_{gro}) = 1033\ (1093)$$

(vi) garnet-biotite geothermometer (Ferry and Spear, 1978)

$$T = (2089 + 0.0096\, P + 1661\; x_{gro})/(0.782 - \log (Bfm/Gfm) + 0.755\, x_{gro}) = 1051\ (1181)$$

(vii) garnet-biotite geothermometer (Loomis and Nimick, 1982)

$$T = (2556 + 0.0100\, P + 1661\, x_{gro})/(1.178 - \log (Bfm/Gfm) + 0.755\, x_{gro}) = 1069\ (1182)$$

(viii) garnet-biotite geothermometer (Thompson, 1976)

$$T = (2731 + 1661\, x_{gro})/(1.548 - \log (Bfm/Gfm) + 0.755\, x_{gro}) = 980\ (1072)$$

(ix) Estimate of water fugacity at 1063 K, 8600 bars from 4 clinozoisite + quartz = 5 anorthite + grossular + 2 water

$$25638 - 104.138\,(T-Tr) + 1.478P - 0.525\,(T-Tr-T \log (T/Tr))$$
$$0.006582\,(T-Tr)^2 + 3.8034\,(T-Tr)^2/T + 2\,RT \ln fw + RT \ln K = 0^*$$

*data from Helgeson and others (1978). Tr = 298.15 up to 848 K where a correction for quartz inversion is added. K = 0.6711

$$2\,RT \ln fw = 34379 \quad fw = 3421 \quad pw = 2584 \text{ bars}$$

(b) Jumping Brook complex (specimen 99-2)

$$\text{garnet; } x_{gro} = 0.352\; x_{and} = 0.597\; x_{cz} = 0.175\; x_{ep} = 0.725\; x_{pr} = 0.85$$

(and = andradite; cz = clinozoisite; ep = epidote; pr = prehnite)

(i) 2 epidote + grossular = 2 clinozoisite + andradite

$$2830 + 0.200T + 0.012P - 13.943\,(T-Tr-T \log (t/Tr)) - 0.011710\,(T-Tr)^2$$
$$-3.300\,(T-Tr)^2/T + RT \ln K = 0^*$$

*data after Helgeson and others (1978). Tr = 298.15. K = 0.098817

For P = 3000 bars, T = 628 K and dT/dP = 2°/kilobar

(ii) Estimate of water fugacity at 628 K and 3000 bars

2 clinozoisite + 2 grossular + 3 quartz + 4 water = 5 prehnite

$$-131622 + 234.6\; T + 2.643\; P - 4\; RT \log fw + RT \log K = 0^*$$

*Data of Liou (1971) fitted to appropriate form. K = 116.931

$$fw = 389 \text{ bars} \qquad Pw = 2664 \text{ bars}$$

(iii) Fe-Mg exchange geothermometry of pelitic rocks

Table 5 (cont'd)

Specimen	19-2	30-1	28-3	10	43-1
biotite-garnet Bfm/Gfm	-	0.132	0.165	0.178	0.235
L-N	-	841	909	929	989
F-S	-	810	885	907	972
T	-	795	852	868	911
chlorite-garnet Cfm/Bfm	0.118	0.122	0.147	0.138	0.234
L-N	843	862	899	907	1623*
garnet-cordierite Cdfm/Gfm	-	-	-	6.188	-
C(T)				951(979)	
P (P-E)	-	3600	4500	8200	5400
P (assumed)	3000	3100	3400	3600	4000
T (assumed)	820	840	895	915	980

Abbreviations: L-N Loomis and Nimick (1982); F-S Ferry and Spear (1978); T Thompson (1976); C Currie (1971); P-E Powell and Evans (1983); Cfm = Fe/Mg in chlorite; Cdfm = Fe/Mg in cordierite. Geothermometers not specified in Table 5(a) are: chlorite-garnet $T = (2745 + 0.0147P + 1661\ x_{gro})/(1.253 - \log\ (Cfm/Gfm) + 0.755\ x_{gro})$ garnet-cordierite (Currie) $T = (4515 + 1661\ x_{gro})/(6.37 - \log\ (Gfm/Cdfm) + 0.755\ x_{gro})$; garnet-cordierite (Thompson) $T = (2724 - 1661\ x_{gro})/(0.884 + \log\ (Gfm/Cdfm) - 0.755\ x_{gro})$.

Peak pressure-temperature conditions can be estimated from a number of lines of evidence. The experimental work of Aitken (1983) suggests that the assemblages containing scapolite + diopside (rather than plagioclase + calcite) equilibrated at temperatures greater than 625°C in the presence of a relatively CO_2-rich fluid. The composition of the observed scapolite (Appendix 3, specimen 54) approaches that of the most stable scapolite found experimentally by Aitken (1983) suggesting that the temperature need not lie much above this value. Such temperatures, say 625-675°C, would be compatible, at moderate water activities, with minor partial melting of the gneiss (Brown and Fyfe, 1970; Winkler, 1979) in accord with limited distribution of migmatite in the Pleasant Bay complex, the widespread occurrence of lenticles of leucocratic quartzofeldspathic material suggestive of local partial melts, and the rather sharp distinction between cross cutting granite gneiss and gneissic host. At such temperatures the observed stability of epidote suggests water pressures in excess of 500 MPa (5 kilobars) (Newton, 1965, 1966), and since water activities are deduced to be at least locally much reduced by admixture of CO_2, the total pressure probably was substantially greater than 500 MPa (5 kilobars). The general aspect of the terrane therefore suggests peak metamorphic conditions in the range of 625-675°C at pressures greater than 500 MPa (5 kilobars) in the presence of a relatively CO_2-rich fluid.

More detailed estimates can be made by applying geothermometers, particularly to specimen 54 (see data in Appendix 3). The potential P-T indicators considered are listed in Table 5. Two of these indicators are believed to be relatively insensitive to pressure, namely the scapolite-plagioclase geothermometer (Goldsmith and Newton, 1977; Otterdoom and Gunter, 1983) and the garnet-hornblende geothermometer (Graham and Powell, 1984). For the mineral compositions in the calcareous bands these geothermometers estimate 1073 and 1064°K respectively, or an average of 1069°. Using this value as an initial estimate, the plagioclase-pyroxene-garnet geobarometer (Perkins and Newton, 1981) can be applied

$$P = 675 + 17.179T + 3.5962T \ln a^2{}_{gr} \cdot a_{py}/a_{di} \cdot a_{an}$$

where a_i refers to the activity of component i, and the subscripts are as follows; di = diopside component in clinopyroxene, an = anorthite component in plagioclase, gr = grossular component in garnet, and py = pyrope component in garnet. These activities, as defined by Perkins and Newton and calculated from the data of Appendix 3 are as follows

$$a_{an} = (x_{an}\ (1 + x_{an})^2/4) \cdot g_{an}$$
$$RT \ln g_{an} = (1 - x_{an})^2 (2075 + 9318 x_{an})$$
$$RT \ln a_{an} = RT \ln 0.618 + 452.2 \text{ (for } An_{78})$$

where g_i is the activity coefficient of i. Perkins and Newton (1981) give a treatment of activities for garnet components based on the "symmetrical solution" treatment of Ganguly (1979), but with differing solution parameters. Labelling the garnet components almandine, pyrope, grossular and spessartite as 1, 2, 3, and 4 respectively, the correction for activity according to Ganguly (1979) can be written

$$RT \ln a_2 = RT \ln x_2 + w12\ x_1{}^2 + w23\ x_3{}^2 + w24\ x_4{}^2$$
$$+ (w21 + w23 - w13) x_1 x_3 + (w21 + w24 - w14) x_1 x_4 + (w23 + w24 - w34) x_3 x_4$$

with symmetrial expressions for the other components. Newton and Perkins (1981) give the solution parameters as w12 = w13 = 0, w23 = 3300-1.5 T(calories), and assume, in effect, w14 = w24 = w34 = 0. Ganguly (1979) gives the parameters w12 = 2979, w13 = 678, w23 = 3280, w14 = 350, w24 = 3200, w34 = 0, noting that these parameters are derived for 640+/-40°C. The difference in parameters is significant. Calculation of P from the equation above gives 6880 bars if the Perkins/Newton parameters are used, but 10339 bars if the Ganguly parameters are used.

Given an estimate of pressure, other geothermometers based on Fe-Mg exchange can be applied, notably the garnet-pyroxene geothermometer (Ellis and Green, 1979; Ganguly, 1979) and the garnet-biotite geothermometer (Thompson, 1976; Ferry and Spear, 1978; Loomis and Nimick, 1982). With the exception of the Ellis/Green calibration, which is based on empirical data, all these calibrations are subject to uncertainties about correction for the composition of garnet outlined above. Assuming a pressure of 8.5 kilobars, the Ellis/Green garnet-pyroxene geothermometer gives a temperature of 1071°C (1066 if 6.8 kilobars is assumed). The other geothermometers turn out to be strongly dependent on the choice of solution parameters for garnet, particularly the garnet-biotite geothermometer which gives results differing by almost 200°C for the two sets of parameters listed above. The parameters listed by Perkins and Newton (1981) produce reasonable agreement among all the geothermometers, except the calibration of Thompson (1976), while the parameters of Ganguly (1979) lead both to disagreement, and to unreasonably high temperatures (compare Table 5a). The parameters of Perkins and Newton (1981) have therefore been adopted. These parameters lead to smaller corrections than those proposed by Pigage and Greenwood (1982) which were based on the parameters of Ganguly (1979). Based on the geothermometry outlined in Table 5, I adopt a temperature of 1063°K, or 790°C with an uncertainty of ±30°. The relative uncertainty in pressure is much larger. I have assumed a pressure of 8.5 kilobars with an uncertainty of ±1.5 kilobars.

The water fugacity can be estimated from consideration of the epidote-anorthite-grossular-water assemblage. An equilibrium can be written among these minerals

4 clinozoisite + quartz = 5 anorthite + grossular + 2 water

Thermodynamic properties for all these components are tabulated by Helgeson and others (1978). Designating the change in standard free energy at P.T calculated by the prescription of Helgeson and others (1978) as dG°, the condition for equilibrium can be written

$$dG^\circ + RT \ln K + 2\,RT \ln f_W = 0$$

where f_W is the fugacity of water, and K the equilibrium coefficient which can be written

$$K = a^5{}_{pl} \cdot a_{gr} / a^4{}_{cz}$$

Expressions have already been given for the activity of plagioclase and grossular. The activity of clinozoisite is presumed to be equal to its mol fraction (0.64) as shown by Bird and Helgeson (1980). These data lead to $2\,RT \ln f_W = 34379$, or $f_W = 3421$ bars. From the tables of Burnham and others (1969), the fugacity coefficient of water at these conditions is 1.324, giving a water pressure of 2584 bars. Note that this estimate depends on the assumption that clinozoisite equilibrated at the same P-T conditions as the other minerals considered.

The quartz-plagioclase-biotite-hornblende gneisses surrounding the calcareous lenses do not lend themselves to accurate estimates of P and T because of lack of knowledge on the activity relations of amphibole components (compare Oba, 1979). However, their water fugacity can be compared with those of the calcareous rocks by consideration of the assemblage hornblende-plagioclase-epidote-quartz. An equilibrium among components of these minerals can be written

4 tremolite + 5 albite + 14 anorthite + 4 water =
5 pargasite + 6 clinozoisite + 27 quartz

As previously shown

$$dG^\circ - 4\,RT \ln f_W + RT \ln K = 0$$

where K is the equilibrium constant. For the present purpose it is convenient to rearrange this expression to

$$f_W = K \exp (G^\circ/RT)/4$$

which shows that at fixed P, T the fugacity of water is directly proportional to K, which can be written as follows;

$$K = a^5{}_{pa} \cdot a^6{}_{cz} / a^{14}{}_{an} \cdot a^5{}_{ab} \cdot a^4{}_{tr}$$

Formulae are given above for activity of plagioclase and clinozoisite. Lacking other information I assume the activity of amphibole components to be equal to their mol fraction, giving

$$a_{tr} = (Ca/2)\,(Si-6)/2.\,(Mg/5)$$
$$a_{pa} = Na.Ca/2.(Al^{vi})/2.(Mg/4).(Al^{iv})$$

These formula cannot be accurate since non-ideality is well established in amphibole (Oba, 1979). For purely comparitive purposes however they permit qualitative assessment of differences in water fugacity. For specimens 54 and 49-1 (Appendix 3) log K is calculated to be 13.165 and 17.049 respectively. Hence the calcareous assemblage had a water fugacity significantly lower than that in the surrounding more granitoid gneisses. (The ratio of water fugacities is exp ((13.165-17.049)/4) or 1/2.64.) The low water fugacities in the calcareous bands, which have granulitic affinities, could be due to local dilution of the metamorphic fluid by CO_2 or other phases.

Similar conclusions may apply to granulite-facies assemblages reported north of the area covered by this report (MacDonald and Smith, 1979). Aitken (1983) noted the common occurrence of local granulite facies assemblages in generally amphibolite-facies terranes.

Field, petrographic and mineral chemistry data suggest that the biotite gneisses of the Pleasant Bay complex equilibrated about 550 Ma at conditions near 790°C and 8.5 kilobars with water pressure a significant fraction of total pressure. Such conditions fall in the high-pressure and high-temperature part of the amphibolite facies (compare Winkler, 1979, p. 65). Under such conditions extensive partial melting of compositionally favourable horizons would be expected producing the observed migmatitic character of the gneisses. Such P-T conditions would be expected only under deep-seated conditions corresponding to depths of 25 km or more.

Metamorphism of the Jumping Brook complex

The Jumping Brook complex consists of three members, namely a volcanic member, a pelitic member and a psammitic member. The complex exhibits strong variation in metamorphic grade so that all three members can be seen at both low and high metamorphic grades. I here consider the three units in order.

Volcanic member (unit Sjbv)

Rocks of apparent volcanic or hypabyssal character were assigned to the Jumping Brook complex if they were schistose, rather than gneissic (in which case they were assigned to the Pleasant Bay complex), or essentially massive to vesicular (in which case they were assigned to the Fisset Brook Formation). The rocks so assigned vary from very low grade schists containing diaspore, chlorite and epidote, to relatively high grade amphibolitic schist containing oligoclase and hornblende. Low grade rocks form a discontinuous fringe along the western side of the highland extending from Pillar Rock north across Trout Brook, and from Chéticamp River south-southeast to Margaree River. The rocks in the northern fringe from Trout Brook to Pillar Rock have been heavily sheared, and now are partly mylonitic (Fig. 27) but their chemical analyses (Appendix 2, analyses 13-2 (compiled in unit Mbr) and 12-3) leave no doubt that they represent igneous rocks or volcanogenic sediments. The mafic material in many examples exhibits fine quartz-rich lenticles probably relict from volcanoclastic sediments. The acid portion, well displayed along the Cabot Trail east of Pillar Rock contains very characteristic sanidine and anorthoclase phenoclasts (Fig. 28) in a finely sheared paste demonstrating that they derive from acid volcanic flows or crystal tuffs. In both acid and mafic portions the matrix consists of finely ground material containing recognizable epidote, brownish fine grained material identified by microprobe and X-ray diffraction as diaspore, albitic plagioclase and quartz. The presence of diaspore + quartz, rather than the chemically equivalent, pyrophyllite, suggests very low grade metamorphism, possibly related to weathering or diagenesis. Since this material occurs in a cataclastic paste, it could conceivably have developed during the latest episode of cataclasis in Mississippian time. However the assemblage (diaspore + epidote + albite) is compatible with low pressure hydrothermal alteration, and apparently primary igneous feldspars are locally preserved. I interpret the observations to indicate very low grade metamorphism prior to cataclastic deformation.

Somewhat higher grade rocks occur along Chéticamp River and to the south along Faribault Brook. Along Chéticamp River recognizable schistose mafic flows outcrop in the stream, badly fractured and strongly veined by epidote and prehnite. To the south the rocks are thinly laminated with alternating 1 to 10 cm bands of salic and mafic material. In thin section the rocks consist of imperfectly oriented aggregates of tremolite, biotite, prehnite and epidote. With the exception of the very lowest grade (mylonitic) material, all the rocks of this unit exhibit a diagnostic texture consisting of ragged rosettes of minerals surrounding, or cutting across, larger, unoriented, ragged porphyroblasts (Fig. 29). Feebly pleochroic garnet (rich in andradite and grossular components) occurs in veins with prehnite and epidote. The more salic parts of the unit contain no tremolite or biotite, but contain relatively large amounts of muscovite, along with quartz, albitic plagioclase and epidote. Prehnite occasionally occurs in veins in the salic bands, but not in the bulk of the rock. Along the margin with the Salmon Pool granite (Dsp) and unit SDg to the south, the pale blue green amphibole develops yellow-green rims. Electron probe analyses (Appendix 3, specimen 99-2) show that this rim amphibole contains significant Al and is

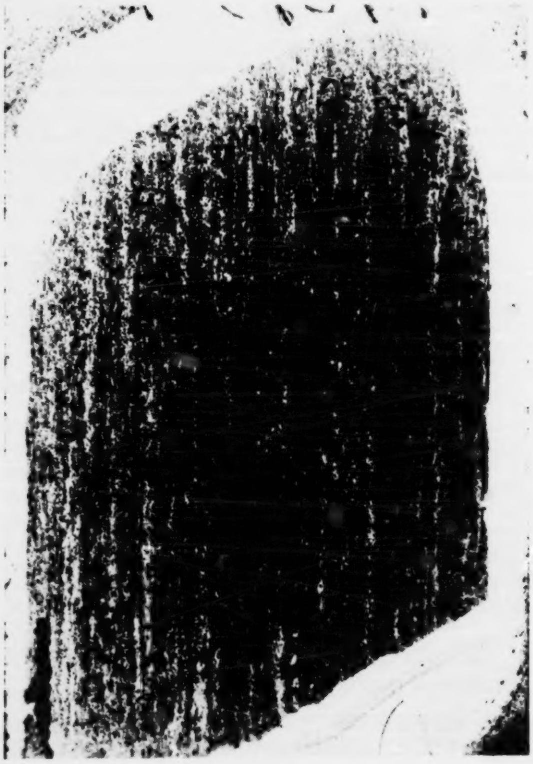


Figure 27. Mylonitic volcanogenic metasediment. Pale streaks are rich in quartz, dark grey material mainly chlorite, black spots are diaspore. Width of field 3 cm. (GSC 202868-T)

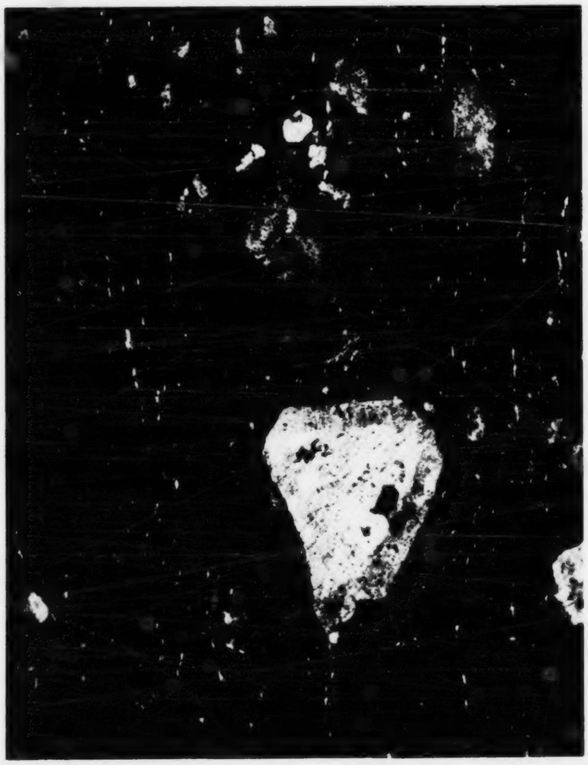


Figure 28. Anorthoclase phenoclasts in fine grained cataclastic paste. Width of field 1 cm. (Specimen 12-3). (GSC 204153-A)

transitional from actinolitic amphibole to hornblende. Correlative rocks east of the Salmon Pool pluton, well exposed at the top of Robert Brook, contain similar assemblages with blue green amphiboles grading to yellow-green amphibole, but contain a more calcic plagioclase (An_{15-20}) and notably less epidote. Prehnite and grossular are absent from these rocks. The highest grade rocks observed in the unit occur near the head of Jumping Brook, and consist of hornblende, biotite, oligoclase to andesine (about An_{30}) and rare pink garnet. According to usual definitions (Turner, 1981; p. 365) these rocks lie within the amphibolite facies, whereas the other hornblende-containing rocks probably lie in epidote amphibolite facies, and the tremolite-biotite-albite assemblages fall in greenschist facies. These assemblages give little clue as to the pressure of metamorphism, but the absence of pumpellyite or lawsonite from the prehnite-bearing assemblages suggests that pressure was not very high.

Chemical analyses of the Jumping Brook volcanics have been compiled in Appendix 2, and compared to similar rocks in Table 3. The average analyses show considerable resemblance to the volcanics of the Pleasant Bay complex, as well as to the average tholeiite of Nockolds (1954) and related rocks. The Trout Brook rocks exhibit statistically different TiO_2, FeO, and Fe/Mg ratios than the Pleasant Bay amphibolites, but the similarities raise the possibility, mentioned by Jamieson and Craw (1983), that the amphibolitic rocks of the Pleasant Bay complex, whether of several ages or not, are higher grade equivalents of the mafic schists of the Jumping Brook complex. This possibility seems unlikely since Pleasant Bay and Jumping Brook rocks can be seen intercalated at several localities, for example along Fishing Cove River and in the upper part of Chéticamp River, and there appear to be quite distinct in terms of metamorphic grade and structural style. The Jumping Brook rocks nowhere exhibit the potash feldspar porphyroblasts characteristic of much of the Pleasant Bay amphibolite. If the two sets of amphibolites are equivalent, then the Chéticamp region exhibits a remarkably complete section through the crust at time of latest metamorphism, since all grades of metamorphism from diaspore-chlorite to granulite are exposed over a distance of about 10 km. If the two sets of amphibolites are distinct, the similarity in composition is not particularly surprising since the composition is that of an abundant and ubiquitous magma type.

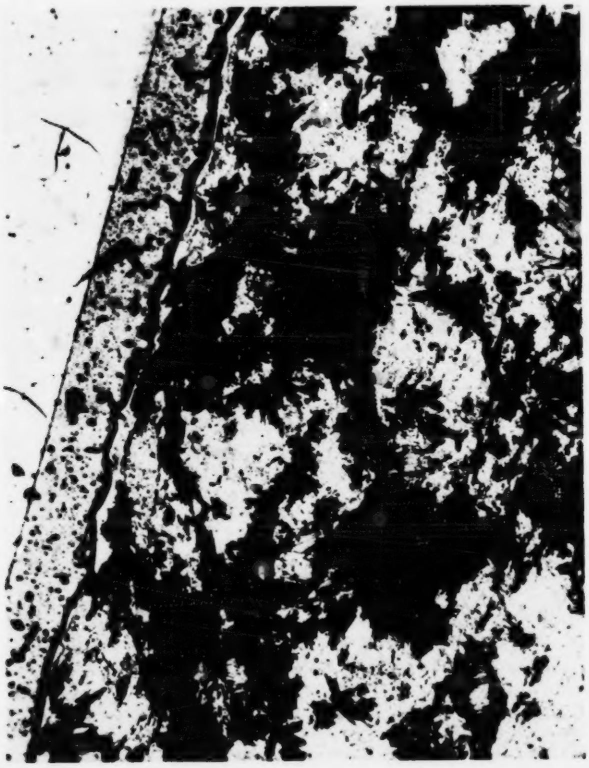


Figure 29. Rosettes of tremolite overgrowing ragged biotite in texture typical of Jumping Brook volcanics. Width of field 1 cm (Specimen 69-1). (GSC 204153)

Pelitic member (unit Sjbp)

Pelitic rocks of the Jumping Brook complex occur interbedded with mafic volcanic rocks in the upper parts of Faribault and Daphinée brooks, and elsewhere form a layer some tens to hundreds of metres thick stratigraphically overlying the volcanic rocks. The pelites can be easily recognized by their very strong, commonly plicated, schistosity. At low metamorphic grade the rocks are black or greenish black, but at a higher grades they take on a distinctive silvery sheen on both fresh and weathered surfaces, commonly fracturing on spectacular nodular surfaces (Fig. 12). Muscovite-rich schists typically occur interbedded with more quartz-rich and less fissile material on the scale of a few centimetres. In many cases the distinctive metamorphic minerals are confined to thin layers within the muscovite schist.

In thin section muscovite dominates the appearance of the rocks. At lower grade muscovite forms an aligned network of large lenticular grains around quartz augen, with some larger lenticles of polycrystalline mosaic quartz. Chlorite occurs sparingly as shreds within muscovite, and granules of yellowish epidote are scattered sparsely through the thin sections. About one third of the thin sections of low metamorphic grade contain sufficient amounts of organic matter that the sections appear very dusty to almost opaque. Attempts to X-ray this material showed it to be amorphous. Some specimens contain up to 2-3 per cent of prismatic, olive-coloured tourmaline.

With increasing metamorphic grade minerals indicative of various grades of metamorphism appear successively, but because of sporadic outcrop and structural complications, no single cross strike series of exposures yields a complete set of index minerals. Chloritoid accompanied by garnet appears in the middle part of Trout Brook and the lower part of Faribault Brook. Chloritoid appears as idioblastic forms commonly with slightly ragged terminations, which cut across the oriented muscovitic matrix (Fig. 11). Accompanying garnet forms small, dense granules not readily visible in hand specimen. The biotite isograd can be observed in Faribault Brook, and north of Jumping Brook. Biotite first appears as small greenish brown idioblastic forms along partings in muscovite. Biotite and chloritoid nowhere coexist. An almost continuous section on Corney Brook exhibits successively the first appearance of staurolite, cordierite and andalusite in that order. The porphyroblasts tend to irregular blobby shapes full of inclusions. Staurolite and cordierite are riddled with rounded quartz and feldspar grains, whereas andalusite, which superficially looks to be slightly augened pink idioblasts on Corney Brook, turns out in thin section to be riddled with inclusions including staurolite and garnet. Along the middle part of Chéticamp River canyon, kyanite occurs together with garnet and staurolite, and in some cases andalusite. The kyanite is invariably in small, corroded grains wrapped in a shell of muscovite. It was nowhere observed in direct contact with garnet, staurolite or andalusite, all three of which occur intergrown with each other. The kyanite is interpreted to be relict.

The distribution of metamorphic minerals as presently known is shown in Figure 2. The metamorphic grade tends to increase from west to east. In relatively low-grade rocks the metamorphic minerals cut the foliation, and appear to be essentially unoriented (Fig. 11), whereas in high-grade rocks the prophyroblasts are wrapped in the foliation, commonly forming augen (Fig. 12). Petrographic and electron probe observations suggest a three stage growth history of the high-grade metamorphic minerals, best exemplified by garnet. The probe data are summarized in Appendix 3. The undeformed garnets in low grade rocks (chloritoid zone, specimen 19-2), and the cores of some higher grade garnets (specimens 10, 28-3) are inclusion free and MnO-rich (up to 8.64%). In staurolite-grade rocks, the bulk of the garnet contains numerous small inclusions (specimen 10, 28-3), which may show rotated or sinuous inclusion trails (Fig. 30), and is distinctly poorer in MnO (5-6%) than the lower grade material. These higher grade garnets commonly exhibit thin, inclusion-free rims which may cause the porphyroblasts to approach idioblastic forms. The latest garnets (specimen 30-1) contain very little MnO (less than 1%), but their Fe/Mg ratios are about the same as those of the inclusion rich material. These observations can be interpreted to suggest initial crystallization of garnet in a static environment subsequent to formation of the aligned muscovite matrix and quartz augen during an early period of deformation. The small size of the mineral cores suggests that growth was slow and conditions barely within the stability field of garnet. As P-T conditions increased (lower MnO content) growth rate of garnet increased (inclusion-rich crystals), and deformation recommenced (rotated inclusion trails). The final growth of garnet occurred after deformation ceased, and Mn had been essentially depleted. Similar three stages histories can be deduced from staurolite and cordierite porphyroblasts (initial dense nuclei, strongly poikilitic central zones, dense rims), although zoning in composition is weak or absent. The lone cordierite analyzed (due to problems with alteration) is relatively rich in Fe (Fe/Fe + Mg = 0.55) suggesting relatively low temperature and pressure.

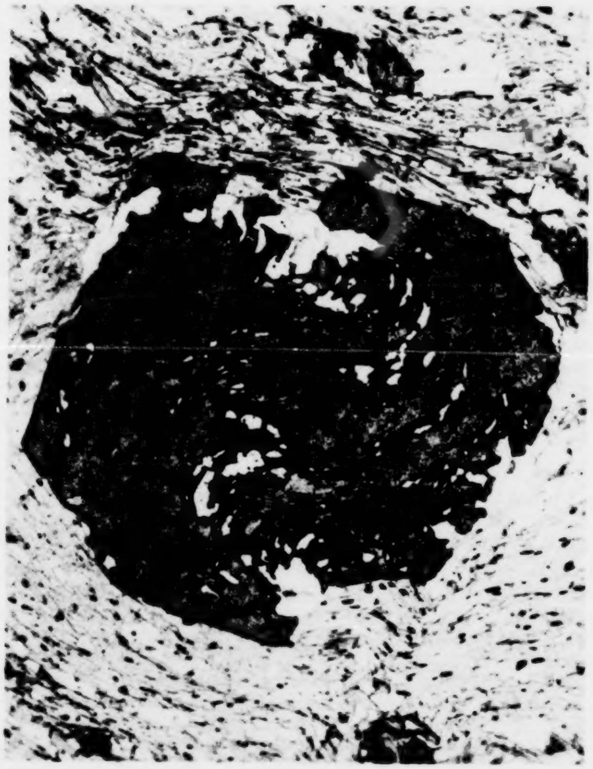


Figure 30. Rotated inclusion trails in garnet porphyroblast. Pelitic schist of the Jumping Brook complex (specimen 74030-1). (GSC 202832-L)

Chlorite appears to be stable in almost all of the specimens examined, commonly occurring intergrown with muscovite. The composition of the chlorite becomes progressively more magnesian with increasing grade (Appendix 3), again suggesting that it is a stable phase. The late stage retrogressive development of chlorite and epidote, ubiquitous in the Pleasant Bay complex, appears in the Jumping Brook complex only in narrow zones along shear planes.

Bulk rock chemical analyses for the modally analyzed specimens of the pelitic schists of the Jumping Brook complex are given in Appendix 2. The analyzed specimens were relatively large, and therefore include both the strongly aluminous layers previously noted as well as more quartzose material. The analyses therefore approach an estimate of the overall composition of unit Hjbp. The analyses are characterized by high K/Na ratios, and low CaO content (average 0.69 weight per cent CaO), characteristics typical for pelitic rocks and expressed modally by the ubiquitous presence of muscovite and the complete absence of plagioclase. Such rocks are not found in the Pleasant Bay complex, and no chemical analyses approaching these characteristics were recorded. Conversely, no calcareous material similar to that found in the Pleasant Bay complex has been recorded in the Jumping Brook complex. The substantial component of pelitic material in the Jumping Brook complex contrasted to the substantial component of calcareous material in the Pleasant Bay complex strongly suggests that the two complexes formed from essentially different protoliths, although the exact boundary between them is now obscure.

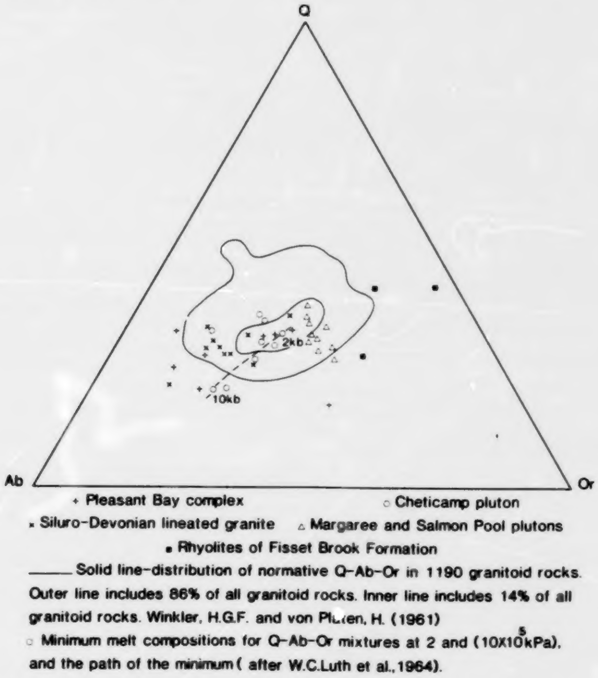


Figure 31. Plot of granite analysis in the Q-Ab-Or system using normative components.

Psammitic member (unit Sjbg)

In hand specimens the greywackes of the Jumping Brook Group are fine grained, homogeneous grey to greenish grey rocks containing sufficient muscovite and/or chlorite to give the rocks a phyllitic sheen. In thin section the rocks are dominated by quartz. In some rocks along the Cabot Trail south of Corney Brook original sedimentary quartz grains can still be recognized, although now severely strained and locally recrystallized to a mosaic of sub-grains. East of Cap Rouge the rock contains small clasts of fine grained quartz thought to be derived from gneiss and schist of the Pleasant Bay complex, and grains of partially sericitized andesine and microcline similar to that found in the Cheticamp pluton. These low grade rocks have a phyllitic appearance due to varying amounts of white mica, chlorite and stilpnomelane.

Slightly higher grade rocks form a belt extending south along Faribault Brook to the southern boundary of the mapped area. The quartz is coarser grained, and clastic textures can no longer be recognized, although local ribbon-like masses of quartz may represent highly deformed clasts. Stilpnomelane is absent from these rocks, and the white mica has high birefringence and good crystal form suggestive of well crystallized muscovite. Biotite and garnet occur as small idioblastic porphyroblasts cutting the foliation. The garnet exhibits high MnO content (Appendix 3) suggesting the biotite-garnet pair is about equivalent in grade to the nearby garnet-chloritoid pair in the pelitic rocks. Garnet is particularly abundant in the greywackes adjacent to pelitic bands. The greywackes quite consistently contain orthoclase and oligoclase to andesine, presumably relict from detrital grains. There is no observable tendency toward formation of albitic plagioclase, even in low grade rocks, nor is there any obvious growth of more ordered potash feldspar.

Higher grade equivalents of the greywackes can be observed along Corney Brook. With increasing grade there is a tendency for the rocks to become markedly less foliated, until they approach a pale grey, almost aplitic texture, only faintly banded, with a vague parting parallel to the banding. The textures approach xenomorphic granular, with a preponderance of three-point grain boundaries. Biotite and muscovite become markedly more equant, and their orientation and concentration in seams becomes less. At about the stage where andalusite appears in the pelitic rocks, chlorite disappears from the greywackes. Since all the greywackes contain two feldspars, these high grade rocks approach in composition foliated two-mica granites with a little garnet.

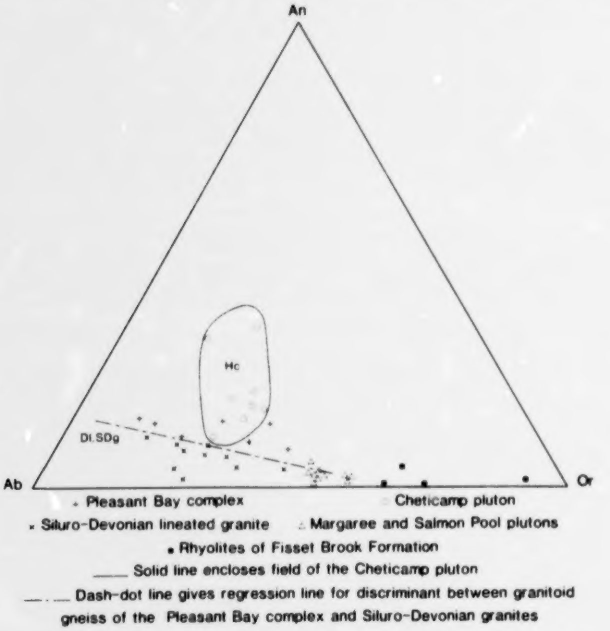


Figure 32. Plot of granite analyses in the Ab-Or-An system using normative components.

Possibly because of their texture, the greywackes were much more subject to regressive metamorphism than the pelitic rocks associated with them. Virtually all the biotite of the high grade greywackes has been partly chloritized, and many of the high grade rocks have films of chlorite between grains. The composition of this secondary chlorite is quite constant and iron-rich (Appendix 3) and does not vary with grade of the relict higher-grade minerals.

In bulk rock chemistry the greywackes form a fairly well defined group, richer in SiO_2 and poorer in aluminum than the pelitic schists. The analyses are in a broad sense granitic (Appendix 2) but the appreciable content of MgO, FeO, Cr and Ni suggests that some component of the sediments was derived from mafic rocks. The analyses overlap those of the biotite gneiss and schist of the Pleasant Bay complex, although there is a noticeable tendency for the Jumping Brook complex to be higher in SiO_2 and lower in total iron. The biotite gneiss of the Pleasant Bay complex (unit **Hpbp**) does not have Zn contents approaching the high values seen in some Jumping Brook rocks. An analysis of variance showed the two units cannot be statistically separated on the basis of bulk rock chemistry at the 1% confidence level, although SiO_2, FeO(t), and Zn are distinct at the 5% confidence level.

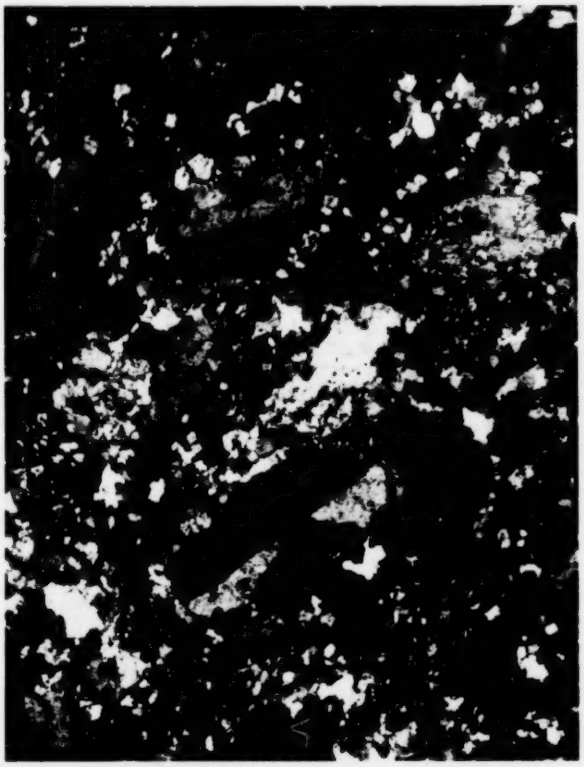


Figure 33. Relict Carlsbad and albite twins in granulated Cheticamp pluton. Width of field 1 cm. (Specimen 41-2). (GSC 204153-B)

Pressure and temperature of metamorphism in the Jumping Brook complex

The Jumping Brook complex, unlike the biotite gneiss of the Pleasant Bay complex, exhibits strong variation in metamorphic grade which can be examined in both pelitic and basaltic compositions. The presence of cordierite and/or andalusite in the high-grade pelitic rocks suggests a low to moderate pressure metamorphic series. The occurrence of early kyanite and late andalusite implies metamorphism must have been due to increase in temperature at almost constant pressure, as can be seen from the aluminosilicate phase diagram (Fig. 8). Assuming that pressure increases with increasing temperature, the maximum pressure must be less than 5.5 kilobars (according to Richardson and others, 1968), or less than 3.8 kilobars if the phase diagram of Holdaway (1971) is preferred. The relatively low pressure-high temperature metamorphism of the Jumping Brook complex is typical of many Acadian contact aureoles in the Canadian Appalachians.

Although P-T contouring of the Jumping Brook complex cannot be carried out because of its complicated fault pattern, dT/dP can be estimated from geothermobarometry of selected specimens. The lowest grade mafic volcanics contain diaspore + epidote + chlorite. The discrepancies and uncertainties in thermodynamic data (Helgeson and others, 1978, p. 110-111) make it unprofitable to attempt quantitative treatment of this assemblage, but experimentally based models of Perkins and others (1979) for the lower stability of diaspore relative to kaolinite imply a temperature in the vicinity of 300°C at pressures below 2 kilobars. At slightly higher metamorphic grade the assemblage tremolite-epidote-prehnite-garnet permits a P-T estimate. Consider the equilibrium

2 epidote + grossular = 2 clinozoisite + andradite

Using thermodynamic data from the tables of Helgeson and others (1978) gives the relation

$$2830 + 0.20T + 0.012P - 13.493\,(T-298.15-T\ \ln(T/298.15))$$

$$-0.011709\,(T-298.15)^2 - 3.300\,(T-298.15)^2/T + RT\ \ln K = 0$$

I assume ideal ionic mixing for epidote (Bird and Helgeson, 1980) and for grossular-andradite garnet which leads to

$$K = (\mathrm{Fe3/Al})_{\mathrm{garnet}} \cdot ((\mathrm{Al-2})/\mathrm{Fe})^2_{\mathrm{epidote}}$$

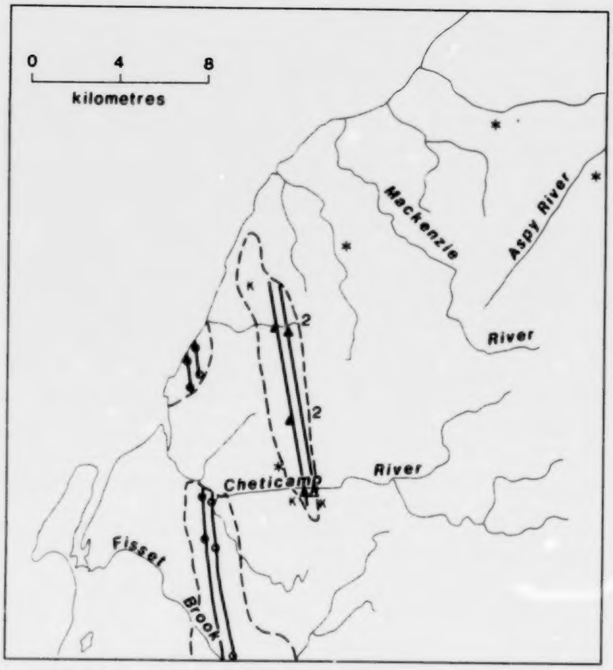
For specimen 99-2, $K = 0.099$. The T dependent part of the above expression changes very slowly with T, giving an estimate of T of approximately $609 + 0.0025P$ or about 615 K for moderate pressure.

A reaction investigated by Liou (1971)

2 zoisite + 2 grossular + 3 quartz + 4 water = 5 prehnite

can be used to estimate the water fugacity in specimen 99-2. The condition for equilibrium can be expressed in the form $a + bT + cP + RT\ \ln f + RT\ \ln K = 0$, where a, b and c are constants to be determined and f is the fugacity of water. The coefficient of P can be estimated from the volume change on reaction (2.643 calories/bar) while a and b can be fitted from the reversed experimental brackets given by Liou (1971) and water fugacity data tabulated by Burnham and others (1969). The result is

$$-131622 + 234.6\ T + 2.643\ P - 4\ RT\ \ln f + RT\ \ln K = 0$$



- - Boundary of Jumping Brook complex
* Granulite facies assemblage in Pleasant Bay complex
● Chloritoid+garnet (no biotite)
○ Biotite+garnet (no chloritoid)
▲ Cordierite (+staurolite+garnet)
△ Andalusite (+cordierite+garnet)
K Kyanite (interpreted to be relict) in Jumping Brook complex

Note that staurolite is widely distributed between the biotite and cordierite isograds, although no isograd has been defined. Jumping Brook complex rocks above andalusite isograd are migmatitic (cf. specimen 43-1)

—— Approximate location of isograd

Figure 34. Approximate position of several metamorphic isograds in the Jumping Brook complex, Chéticamp region.

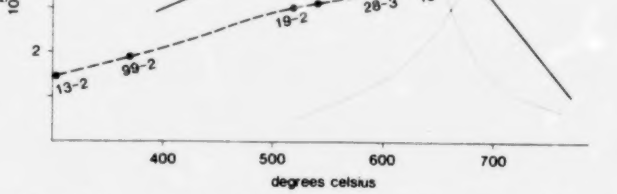


Figure 35. Calculated pressure-temperature conditions in the Chéticamp region. Alumino silicate diagram from Richardson et al. (1968).

Assuming ideal ionic solution for garnet, epidote, and prehnite gives, for the data in Appendix 3, specimen 99-2, $K = 4.7616$. Using the temperature of 615 K previously estimated, this expression predicts a water fugacity of 225 bars.

At higher metamorphic grades, pelitic assemblages give useful information on metamorphic conditions. Richardson (1968) and Rao and Johannes (1979) showed that for iron end members chloritoid + quartz broke down to cordierite-bearing assemblages below 3 kilobars. Since staurolite appears at lower grade than cordierite in the Jumping Brook complex, pressures probably were above 3 kilobars at the staurolite-in isograd. Richardson (1968) also found the breakdown of staurolite + almandine to cordierite to have a very small, positive dP/dT, requiring the pressure for this breakdown to fall in the range 3.0 to 3.7 kilobars. Trial calculations assuming ideal ionic solid solution for garnet, staurolite and cordierite suggest this pressure would be little altered by compositional differences between the analyzed minerals of the Jumping Brook complex and the pure iron end members investigated by Richardson (1968).

Geothermometers based on coexistence of biotite, chlorite and garnet can be applied to pelitic rocks above staurolite grade. These temperature estimates all depend on exchange reactions, and have the form $a + bT + cP + RT \ln K$, where a, b and c are constants and $K = (Fe/Mg)_A/(Fe/Mg)_B$, where A and B are the phases involved. Table 5 exhibits the results of computations using the biotite-garnet and chlorite-garnet geothermometers as calibrated by Ferry and Spear (1978), Thompson (1976) and Loomis and Nimick (1982), as well as results from garnet-cordierite pairs (Currie, 1971; Thompson, 1976). A pressure of 4000 bars was assumed where required, but the pressure correction is less than 5 degrees/kilobar in all cases. All the geothermometers correctly sort the specimens according to field observations, and the pairs (Currie, 1971; Thompson, 1976). A pressure of 4000 bars was assumed where required, but the pressure correction is less than 5 degrees/kilobar in all cases. All the geothermometers correctly sort the specimens according to field observations, and the agreement between geothermometers and between calibrations is reasonable. The calibrations of Ferry and Spear (1978) and Loomis and Nimick (1982) give temperatures significantly higher than that of Thompson (1976), while garnet-cordierite pairs give the highest temperatures. This may be due to slight chloritization of the cordierite, resulting in excessive analyzed Mg values.

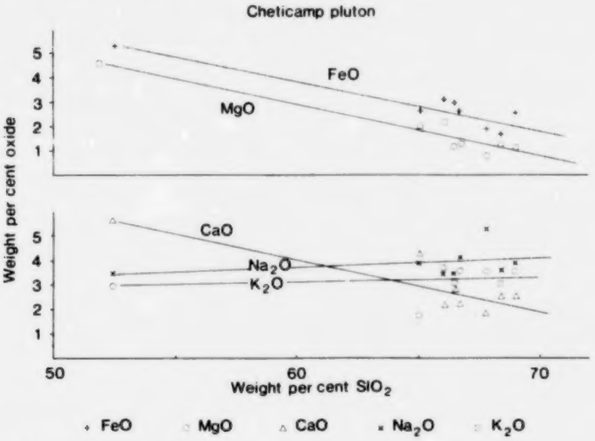


Figure 36. Variation diagram against silica for the Cheticamp pluton. Solid lines are calculated regression lines.

The highest grade pelitic rocks (specimen 43-1 of Appendix 3) contain locally transgressive, fine-grained muscovite granite lenticles interpreted as anatectic melt. Stability of muscovite in contact with a melt requires temperatures of the order of 650-700°C and water pressures of several kilobars. Anderson and Rowley (1981) suggested 2.5 kilobars as a minimum water pressure. Regardless of the exact figure, presence of muscovite at this grade requires water pressures that are a large fraction of total pressure.

Pressure estimates for these rocks can be made from the chlorite-muscovite-biotite geobarometer of Powell and Evans (1984). These estimates give maximum values since the calibration assumes water pressure equal to total pressure. As can be seen from Table 5(b) these estimates are generally similar to those already put forward, although slightly higher, as could be expected. Assuming that the discrepancy results from dilution of the fluid phase and the previous estimates are correct, the partial pressures of water would be 75-85 percent of total pressure.

Pressure-temperature estimates based on Table 5 and the above considerations are shown graphically in Figure 35. These estimates define a P-T trend for metamorphism roughly compatible with the relation $5T = P + 1000$, which implies a dT/dP of 0.2°/bar, or about 60°/km of depth, or at least twice the normal geothermal gradient. This gradient is shown in Figure 35, together with some experimental data. It can readily be seen that rocks in the chloritoid and lower zones lie in the stability field of kyanite, whereas rocks at staurolite grade lie in the andalusite field according to the phase diagram of Richardson et al. (1968), but in the sillimanite field according to Holdaway (1971). Since andalusite-bearing granites and pegmatites are a characteristic feature of the Canadian Appalachians (Currie and Pajari, 1981; Clarke et al., 1976) the diagram of Richardson et al. (1968) appears to be the most applicable. The figure clearly shows why kyanite is occasionally observed as relicts in the lower part of the staurolite zone, and andalusite at higher grade conditions.

Distribution of metamorphic facies

As previously discussed, the rocks of the Jumping Brook complex exhibit a variety of grades of metamorphism. Although the structural complexities are such that continuous isograds cannot be readily drawn, it is apparent that the grade increases rapidly inland. The Pleasant Bay complex, by contrast appears to exhibit essentially the same metamorphic grade over its whole extent, characterized by the assemblage plagioclase-hornblende-epidote, with small zones of granulite facies rocks distributed throughout. This pattern of distribution could be the result of several mechanisms. Most simply it might be supposed that the Pleasant Bay complex represents an older terrane metamophosed in Precambrian(?) time, unconformably overlain by a younger sequence metamorphosed in the Devonian. In this case the P-T progression in the Jumping Brook complex corresponds to exposure of increasingly deeper levels, with the Pleasant Bay complex representing exposed basement, with high grade rocks preserved because they were relatively massive and dry, that is very difficult to re-equilibrate at the lower grade, later conditions. In this case it can readily be shown that no major tectonic dislocations are necessary within the section. The metamorphic conditions within the Jumping Brook complex rise from about 300°C/200 MPa (2000 bars) in the prehnite-epidote facies to about 700°C/400 MPa at the highest grade. The distance between these exposures is about 7 km. This thickness of crust, assuming a mean density of 2.85 g/cm^3, would represent a pressure difference of about 200 MPa (2 kilobars), or that deduced from the field. The thermal gradient would be about 55°/km which is high but comparable to that found in active areas of granitoid emplacement, for example the Himalayas. If it is assumed that the originally horizontal layers are now standing approximately vertical, the relations can therefore be explained without any major tectonic breaks in section. This assumption would be compatible with the steep dips of the Jumping Brook complex. On the other hand, if it is assumed that the Pleasant Bay complex is a more highly metamorphosed equivalent of the Jumping Brook complex, as proposed by Jamieson and Craw (1983), there must be major tectonic telescoping of the section. Since the jump in pressure from the Jumping Brook complex to the Pleasant Bay complex is about 400-500 MPa (4-5 kilobars), this jump would require a vertical displacement between the two complexes of 12-15 km, not to mention any associated horizontal displacement. A yet further alternative would be to abandon the proposed correlation of the Jumping Brook complex and the Pleasant Bay complex, and to assume a major tectonic break between them, as indicated by the great multiplicity of breccia and shear zones. A major objection to models with large tectonic displacements between the Jumping Brook and Pleasant Bay complexes lies in the interleaving of the two complexes along Fishing Cove River and Chéticamp River. This could be tectonic interleaving as indicated by strong shearing around these occurrences, but in this case it would be expected that the very large displacements required would occur along the major shear and breccia zones following the shore of the Gulf of Saint Lawrence, rather than along subsidiary and relatively minor zones lying at an angle to the largest breaks.

It will be recalled that the overall composition of the Pleasant Bay complex differs markedly from that of the Jumping Brook complex. The Pleasant Bay complex contains a significant component of calcareous rocks, and the volcanic material within it (as opposed to later dykes) is of dacitic to andesitic composition. The Pleasant Bay complex lacks strongly pelitic beds. The Jumping Brook complex, by contrast, contains bimodal basalt-rhyolite volcanics, an abundance of pelitic rocks, and no calcareous rocks. In addition, geothermometry shows that the Jumping Brook complex is of moderate pressure-high temperature type, common in the "Acadian" metamorphic rocks of the Canadian Appalachians, whereas the Pleasant Bay complex displays a high pressure-high temperature metamorphism transitional to granulite facies which is unusual, if not unique, in the Canadian Appalachians. These differences suggest that attempts to correlate the two sequences of rocks are liable to confuse, rather than clarify the geological relations. The exact nature of the contact between the two sequences is not clear, but no major displacement has yet been demonstrated, and the relations are compatible with a metamorphosed, moderately deformed unconformity.

MAGMATISM

The Chéticamp region exhibits at least three ages of granitoid emplacement, namely the gneissic granites of the Pleasant Bay complex emplaced prior to the Cheticamp pluton, the Cheticamp pluton itself, and younger granitic rocks probably emplaced in Silurian and later time. Each of these suites of rocks contains younger mafic intrusions, namely the old, porphyroblastic amphibolites in the gneissic granites of the Pleasant Bay complex, schistose amphibolite probably related to the Jumping Brook complex in the Cheticamp pluton, and the Fisset Brook Formation, which locally cuts the younger granites. The granitoid gneisses of the Pleasant Bay complex and their contained amphibolites now have the character of metamorphic rocks, and were discussed in the section on metamorphism. It remains to consider the other magmatic rocks.

Cheticamp pluton

The Cheticamp pluton (Hc) forms an elongate mass extending southerly from Corney Brook to the edge of the mapped area. Over much of this distance the pluton forms the crest of the scarp marking the edge of the Cape Breton Highlands. Smaller occurrences of the same lithology are found within the Pleasant Bay complex, notably as a large dyke-like mass extending from the upper part of Mackenzie River to McIntosh Brook.

Outcrops of the Cheticamp pluton typically exhibit a salmon pink to reddish colour, and brecciated to contorted texture (Fig. 13). The rocks vary from relatively leucocratic hornblende granodiorite to mesocratic biotite granite. The rocks consistently exhibit brecciation and deformation of apparently plastic character. In the least affected rocks, hornblende is converted to chlorite, biotite is iron stained, and the feldspars are cracked. In more severely affected rocks the salic minerals occur in ellipsoidal, contorted pods surrounded by a network of biotite and greenish chloritized material. Along Corney Brook this latter type of material grades to mylonites and flaser gneisses (unit **Mbr**). The least deformed material, exposed in the upper parts of Robert, Fiset and Farm brooks, tends to be distinctly porphyritic. On Farm Brook, microcline megacrysts up to 5 cm in diameter occur. The megacrysts are markedly rounded, sometimes surrounded by granulated haloes ("mortar"), and set in a stained, deformed matrix.

Microscopically the rocks of the Cheticamp pluton exhibit medium to coarse grained inequigranular texture variously modified by cataclasis. Quartz occurs as large isolated grains which may have sutured internal boundaries and a halo of "mortar". Locally the quartz displays an unusual frond-like pattern of extinction, presumably due to some combination of strain and recrystallization. Potash feldspar in this unit commonly exhibits relict Carlsbad twinning (Fig. 33), in many cases with micro-offsets of the twin boundaries. Much of the feldspar appears bluish and chatoyant, with large optic angle but no visible microcline twinning. Irregular patches of both microcline-twinned and albite-twinned material occur within the relict Carlsbad twins, and X-ray diffraction of powdered megacrysts suggests that the material is a mixture of albite and microcline. These properties suggest an originally high-temperature feldspar now re-equilibrated to lower temperature conditions. Plagioclase forms clumps of squarish subhedral grains, dominating the texture of some rocks so that other minerals are interstitial. The twinning pattern is quite complex with blotchy patches of albite twinning combined with pericline twins. Cores, commonly sericitized, range from An_{31} to An_{40}, whereas fresher material in the mantles ranges from An_{16} to An_{22}. The latter composition occurs as myrmekitic intergrowths with potash feldspar ranging from Or_{57} to Or_{80}, with the more sodic compositions dominant. In a few rocks, rims of virtually pure albite overgrow all other feldspar.

The Cheticamp pluton originally contained hornblende and biotite, but these minerals are now largely altered to chlorite, epidote and opaque minerals as a result of cataclastic deformation and retrograde alteration. Pseudomorphs of amphibole occur commonly, but only one specimen was found to contain relict amphibole, in this case a feebly pleochroic paragastic hornblende similar to those in the surrounding amphibolites. Biotite occurs as sheaves of crystals, but the pleochroism is commonly weak, the colour muddy brown, and the material is transitional to chlorite, as confirmed by electron microprobe reconnaisance. In addition to replacing mafic minerals, chlorite also forms films along fractures. Chlorite compositions resemble those in the Pleasant Bay complex (Appendix 3). Epidote of two generations occurs in the pluton. Older epidote is yellow in plane polarized light and forms chains of rounded grains. A younger, almost colourless epidote forms large euhedral prisms around the boundaries of mafic clots. It appears to be an alteration product of amphibole. Electron microprobe analysis suggests a very limited range in composition of the epidote (Appendix 3). Magnetite and hematite occur in the form of a fine opaque dust scattered through the thin sections, or associated with mafic clots and chloritized fractures. None of the opaque minerals is clearly of primary origin. Accessory minerals in the pluton are zircon, apatite, and rare tourmaline.

Modes and chemical analyses of eight selected specimens of the Cheticamp pluton are presented in Appendices 1 and 2. Modally the samples appear to vary from dioritic to quartz monzonitic, although the pervasive alteration makes precise nomenclature difficult. Assuming that the alteration was isochemical, analysis of the chemical data should be more revealing. Plotting variation diagrams (Fig. 35) does not give particularly well defined trends, but the samples share a number of characteristics. Na_2O and K_2O contents show little variation with SiO_2, and Na/K ratios vary relatively little from 1.8, but the amount of CaO decreases markedly with increasing SiO_2 so that the normative plagioclase becomes progressively more albitic. The Fe/Mg ratio is quite remarkably constant around 0.7. All but one of the Cheticamp pluton samples contain significant amounts of Ni and a majority also contain significant Cr. Only the dioritic analysis returns normative diopside. The other analyses contain amounts of normative corundum greater than 1 mol per cent. However the widespread occurrence of hornblende, sphene and magnetite, and the relatively wide range in SiO_2, suggest that the Cheticamp pluton is an I type granite derived from more mafic material. The initial strontium isotope ratio obtained by Cormier (1972) of 0.706 ± 0.004, although rather imprecise, is compatible with direct derivation from mantle material. Normative corundum probably appears due to extensive alteration of the feldspar.

The Cheticamp pluton has clearly been strongly deformed by cataclasis, and from field relations and age it appears to predate the metamorphosed rocks of the Jumping Brook complex. The minerals of the pluton therefore might be expected to exhibit a complex history of metamorphism and deformation. Since the mafic minerals are mainly altered, and nowhere clearly primary, only the salic minerals can be used to estimate the conditions of emplacement and metamorphism. The feldspars are quite complex, so that assumptions are necessary to apply the two-feldspar geothermometer. I assume (a) the Carlsbad twinned feldspar is relict from the original monoclinic magmatic feldspar, and was in equilibrium with the plagioclase phenoclasts, and that the average composition of the Carlsbad twinned material represents the igneous alkali feldspar, and (b) the mymekites represent either late magmatic conditions or possibly recrystallization during metamorphism. On these assumptions the geothermometer returns two values, one representing emplacement temperature and the other either late magmatic conditions or the conditions of metamorphism. For the Carlsbad-twinned material, the correct formula to use is that of Stormer (1975) for disordered material. Assuming an emplacement pressure of 3 kilobars the phenoclasts (Or_{60} - An_{38}) gives values of about 915°C (1188°K), while the smaller grains of feldspar surrounding the large twinned grains (Or_{60} - An_{22}) give 760°C. These temperatures seem quite reasonable for igneous granites, but much higher than those obtained for metamorphism. However the interstitial myrmekite, using the Whitney and Stormer (1977) geothermometer for ordered potash feldspar, gives temperatures of 565°C (Or_{80} - An_{16}). This temperature is similar to those deduced for metamorphism of the surrounding Jumping Brook complex, and thus supports the view that the Cheticamp pluton exhibits relict igneous texture, but was metamorphosed along with the Jumping Brook complex.

The Cheticamp pluton can be distinguished from other granitoid rocks of the Chéticamp region on normative Q-Ab-Or and An-Ab-Or plots (Fig. 31, 32). On both plots there is a notable tendency for the analysis to plot on lines radiating from Q and An respectively, suggesting that some of the variation in composition may be due to fractionation of quartz and calcic plagioclase. This tendency is particularly clear on the An-Ab-Or diagram when contrasted to the chemically similar Pleasant Bay granitoid gneisses. The latter tend to vary strongly in Ab/Or ratios, but relatively weakly in An content, whereas the Cheticamp pluton exhibits relatively constant Ab/Or, but varies in An content.

Muscovite granite gneiss and lineated granite
(units SDg and Dl)

Distinctive leucocratic muscovite-bearing granitic rocks, either gneissic or lineated, tend to fringe the Jumping Brook complex. As noted previously, these rocks locally appear to show gradation both to the Jumping Brook complex and to migmatites of the Pleasant Bay complex. In most cases they occur as myriads of small dykes, or pod-like masses. In thin section these rocks contain large, angular, moderately strained quartz grains lacking mosaic recrystallization. Quartz and potash feldspar form large areas of granophyric intergrowth, analagous to spectacular graphic

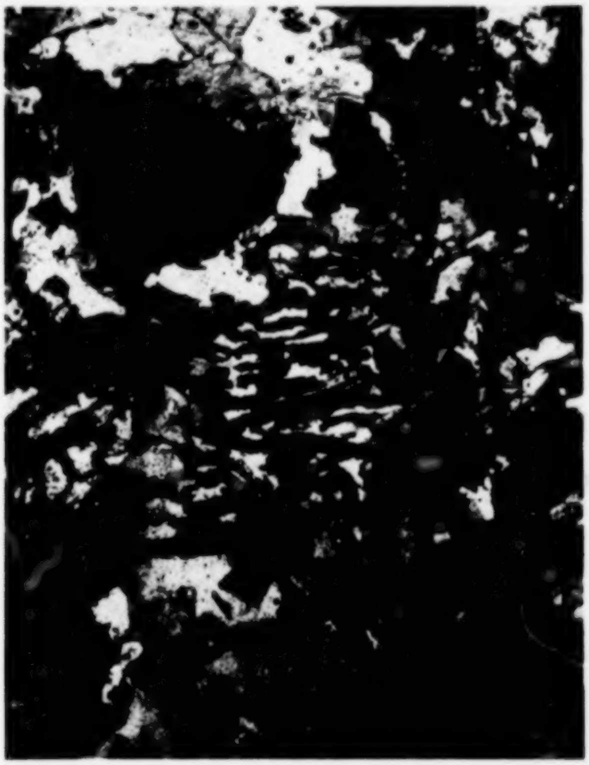


Figure 37. Graphic structure in muscovite granite gneiss. Width of field 50 mm. (Specimen 32-2). (GSC 204153-C)

structures (Fig. 37) locally visible in hand specimen. Much of the potash feldspar exhibits microcline twinning, but internal patches, apparently relict, commonly exhibit a pale blue extinction colour and low optic angle suggestive of monoclinic feldspar. Electron microprobe analyses show this material to lie in the range Or_{84-90} whereas the matrix of the surrounding perthitic microcline is about Or_{93} with exsolved albite (An_3). Plagioclase occurs in subhedral grains zoned from cores of An_{31} to rims of about An_{19}, with occasional thin shells of albite. Blebs of quartz commonly occur in the plagioclase in textures suggestive of perthite-like exsolution. The cores of plagioclase generally show moderate sericitization. Biotite occurs as small euhedral grains concentrated in certain zones to give a rather diffuse foliation. Fresh biotite has a markedly reddish pleochrism, but much of the material has been partially converted to greenish chlorite, either as rims, or along cleavages. Muscovite is much less prominent in thin section than in hand specimens. The rock tends to break along planes containing muscovite flakes, but such flakes are quite sparse in the interior of the broken blocks. The mineral occurs in two generations, either as small flakes similar in morphology to biotite, commonly found as inclusions in feldspar, or as relatively large grains cutting across other minerals and locally forming radiating sheaves. Small euhedral brownish garnets occur in trace amounts in this unit, usually associated with micas, but other accessory minerals such as zircon and apatite appear to be totally absent.

The distinction between foliated and lineated varieties, easy to make in hand specimen, is not obvious in thin section. The lineated rocks appear to be strongly cataclastic. Feldspar grains commonly exhibit a halo of finely granulated material (mortar), suggesting that the spindle-shaped aggregates which give the rock its lineation may be due to cataclasis.

Ten chemical analyses of these granites are compiled in Appendix 2. They show a very limited range of strongly silicic, soda- dominant compositions. The extreme poverty in Ti, P, Cr, Ni, V and Zr is obvious from the analyses. When plotted on normative Q-Ab-Or and An-Ab-Or plots (Fig. 31, 32), the analyses overlap the field of the Pleasant Bay granite gneisses and Cheticamp pluton in Q-Ab-Or, but fall in a distinctive low An position on An-Ab-Or. Like the Pleasant Bay rocks they show a broad range of Ab/Or values, but they clearly fall in a more Ab-rich, An-poor field than the latter. Rather surprisingly, considering the obvious muscovite and accessory garnet in the rocks, the analyses exhibit relatively low contents of normative corundum. In fact analysis 34-1 contains 0.34 mol per cent normative diopside. The average for five foliated granites is 1.75 mol per cent normative corundum, and for five lineated granites 0.44 mol per cent corundum. As pointed out by Currie and Pajari (1981), these low contents of normative corundum are quite compatible with experimental data on anatexis of pelitic rocks. Incorporation of higher amounts of normative corundum requires metasomatic processes which might broadly be termed hydrothermal.

Some of the rocks of this unit contain the assemblage garnet-muscovite-biotite-potash feldspar-quartz, which is appropriate for geothermometry. However the garnet (Appendix 3, specimen 24-2) is very rich in spessartite component, suggesting that it is unlikely to be of magmatic origin. Calculation of temperature using the biotite-garnet geothermometer (calibration of Ferry and Spear, 1978) gives a K = 6.05, and an estimated temperature (at an assumed pressure of 400 MPa or 4 kilobars) of 785°K (502°C). This temperature clearly cannot refer to igneous crystallization temperatures, but suggests that the garnet is a xenocryst from the nearby Jumping Brook complex. The feldspar geothermometer was also applied to this unit, with the previous assumptions that the most calcic plagioclase was originally in equilibrium with the most sodic alkali feldspar. The results, compiled in Table 4, lie in the range 615-650°C for an assumed pressure of 400 MPa (4 kilobars). These temperatures would be comparible with anatexis, at least for water saturated minimum melting compositions, but seem markedly low for emplacement of water-undersaturated granitic melts from depth. They might result either from metamorphic sub-solidus processes, or generation of minimum melt essentially in situ. The latter process appears in concord with all the field data relating to these bodies.

Margaree and Salmon Pool granites

The youngest granitic rocks in the Chéticamp region are essentially postmetamorphic, but locally show effects of retrograde metamorphism due to cataclastic deformation. The Margaree pluton forms a spectacularly megacrystic mass which clearly truncates the cataclastic zone following the Aspy River valley, but is itself cataclastically deformed along its northeastern edge, and at its southern terminus. The Salmon Pool pluton forms a body of much finer grain, locally porphyritic, but also variably cataclastically deformed, with the southern terminus formed of strongly cataclastically foliated rocks.

The megacrystic feldspar of the Margaree pluton commonly displays Carlsbad twins in thin section, with the boundaries offset along microfaults. Much of the feldspar consists of fine flame perthite with patchy tartan twinning, but large blocky patches or blebs up to 0.3 mm across exhibit chess board twinning. These patches appear to be crystallographically controlled both in geometrical and twin orientation, and may represent exsolution plates. Such zones also form the pale buff rims to the megacrysts which can be readily seen in hand specimen. Flames in perthite, and independent small grains of plagioclase have compositions varying from An_{26} to An_{31}, while the albite-twinned material is essentially pure albite (An_1 or less). The potash felsdpar appears to be of reasonably uniform composition (Or_{89-92}) regardless of mode of occurrence. Quartz in the Margaree pluton appears to have originally formed large subhedral prisms, some of which can still be seen in hand specimen. Microscopically this material has all been converted to masses of small interlocking grains. In hand specimen, the mafic mineral appears to be lustrous black biotite, but in section much of this material has feeble brownish pleochroism, weak birefringence, and appears transitional to chlorite, as verified by electron probe analysis. In some specimens the mafic minerals have been entirely converted to a jade-green, non-pleochroic chlorite, which always accompanies moderately to strongly sericitized, reddened feldspar. Apatite, zircon and magnetite occur commonly as accessories.

The Salmon Pool granite tends to be finer grained than the Margaree pluton, but the texture varies considerably. Potash feldspar grains tend to be larger than those of other minerals, and exhibit patches of braid perthite and albite twinned material thought to indicate exsolution of albite from originally sodic, high temperature feldspar. Plagioclase displays strong zoned extinction, but only patchy and local albite twinning. Optical and electron microprobe determinations suggest that the anorthite content ranges from about An_{32-18}. Myrmekitic intergrowths between plagioclase and alkali feldspar occur ubiquitously, and granophyric to graphic intergrowths of quartz and potash feldspar also occur commonly. The remarkable abundance and variety of intergrowths suggests extensive cotectic crystallization.

The Salmon Pool pluton contains biotite as mafic mineral, and much of the pluton also contains small amounts of muscovite. The biotite has a thin platy habit, and the strong red pleochroism typical of titanian biotites. Much of the mineral has been altered to chlorite, with the unusual development of muddy pleochroism and weak birefringence. Where muscovite is present, virtually all the biotite has been chloritized. Muscovite in this pluton commonly contains numerous small quartz blebs within the mica. Biotite commonly contains the accessory minerals apatite, zircon and magnetite as small inclusions.

Ten chemical analyses of the Margaree and Salmon Pool plutons are compiled in Appendix 2. Like the foliated granites of units **SDg** and **Dl**, they are siliceous (SiO_2 greater than 70%), but unlike these rocks they are markedly potassic, and contain detectable amounts of Ti, P, Ni, V and Zr. On Q-Ab-Or and An-Or-Ab diagrams (Fig. 31, 32) the plutons plot together in a distinctive Or-rich and An-poor position. Like the Cheticamp pluton, they show a strong tendency to a linear spread of the analyses on these plots (constant K/Na ratio). The reason could possibly be some fractionation of quartz.

The field relations of the Margaree and Salmon Pool plutons leave no doubt that they are intrusive in their present position and derived from a source region at greater depth (numerous dykes of Margaree pluton, presence of digested inclusions not referable to the country rocks, intrusive breccias around the Salmon Pool pluton). As previously noted on petrographic grounds, and subsequently supported on chemical grounds, there seems evidence that magmas related to these plutons may have been the source of the rhyolitic part of the Fisset Brook Formation. Nevertheless it is extremely difficult to demonstrate magmatic temperatures in these plutons. The mafic minerals are generally altered, and the feldspars exhibit such complex intergrowths combined with sericitization, reddening and alteration that the feldspar geothermometer seems unlikely to record magmatic events. An attempt to apply this geothermometer to the Salmon Pool pluton in the usual manner (most sodic potash feldspar and most calcic plagioclase) led to an estimate of about 610°C (Table 4), which seems clearly too low for an intruded magmatic rock. Presumably such temperatures in this case, and possibly also in the case of the lineated granites, must refer to post-intrusive recrystalization.

Fisset Brook Formation

The Fisset Brook Formation consists of salic to mafic volcanic rocks and associated dykes exposed in two parallel belts fringing the Cheticamp pluton. The salic rocks appear mainly in Au Coin and Factory brooks. In Au Coin Brook the lower part of the section contains dense, purple flow-banded rhyolite which passes upward into moderately welded tuff or ash fall deposits containing flame-like structures (fiamme) as well as partially resorbed quartz phenocrysts of square cross section, indicative of high quartz. The upper part of this member is vesicular, weakly welded material similar to pumice. This member is capped by a 5 m thick basaltic flow overlain by another 10 m of moderately welded tuff. Similar, possibly correlative, sections occur in Farm and Factory brooks, but in Fisset Brook, only 150 m from Au Coin Brook, rhyolitic rocks are not exposed.

The mafic volcanics of the Fisset Brook Formation consist microscopically of a net of skeletal plagioclase microlites with compositions ranging from $An_{45}$to An_{68}. A few larger plagioclase phenocrysts exhibit hopper textures and partial resorption. Sparse relicts of pyroxene locally exhibit pale pinkish pleochrism, small 2V and bluish extinction typical of titanaugite. Most of the pyroxene has been strongly altered to a pale green non-pleochroic chlorite and opaque dust. Apatite and sphene occur commonly in accessory amounts.

The acid volcanic rocks exhibit microscopically a very fine mosaic texture indicative of complete recrystallization, possibly from glass. These mosaic patches truncate the prominent flow banding. A few specimens retain relicts of orbicular texture. Quartz and feldspar crystals are strongly rounded and corroded, and the flow texture deflects around them. Although some euhedral quartz crystals are present with diamond-like sections indicative of high quartz, most large quartz crystals are rounded and may be xenocrystic. Feldspar crystals display a variety of twins, including patchy albite twinning, but the most common form is well developed microcline grid twinning. Optic angles and degree of ordering as determined by X-ray appear to be consistently high.

Appendix 2 presents 10 chemical analyses of the Fisset Brook Formation, divided sub-equally between the salic volcanics and the mafic volcanics. These analyses fall into no simple pattern. Among the mafic analyses, three contain normative nepheline (11-2, 26-2, 80-2), one contains both normative olivine and normative corundum (89), apparently due to metasomatic loss of CaO, and the other two contain normative quartz. In a general way the analyses resemble those published by Keppie and Dostal (1979) which they considered typical for basalt associations connected with continental rifting. However the presence of corundum in the norms suggests that the pervasive hydrothermal alteration was accompanied by a significant amount of metasomatism. Analysis 18-1 resembles tholeiitic basalt, while analysis 38-1 might be considered an andesitic basalt, but considering the strong evidence for alteration, it seems unlikely that any of the analyses should be considered as andesites in sense advocated by Kelley and MacKasey (1965).

The salic rocks superficially chemically resemble those of the Margaree and Cheticamp pluton, but on plots of normative Q-Ab-Or and An-Ab-Or they fall into a distinct field, which lies outside the common field for granitoid igneous rocks (Fig. 31, 32). The K/Na ratios of these volcanics are so high that metasomatism must have intervened during their formation. The anomalous occurrence of microcline in these rocks rather than a high temperature potash feldspar suggests that this metasomatism may have taken place after consolidation, and be associated with the obvious recrystallization of the rocks.

Summary of the igneous history of the Chéticamp region

The Chéticamp region appears to exhibit several cycles of magmatic activity, each passing from early, large scale acid plutonism to later small scale hypabyssal bimodal activity. All plutonic activity took place within an essentially continental environment.

The migmatitic rocks of the Pleasant Bay complex, which contain nebulous boudined remants of mafic dykes overprinted by one or more generations of feldspar porphyroblasts, may represent the oldest cycle. Much larger, more homogeneous masses of granite gneiss cut the migmatite, but themselves contain boudined amphibolite dykes of tholeiitic chemistry, presumbly derived from diabase dykes. The host granites exhibit normative corundum, modal muscovite and ilmenite, and a narrow range of SiO_2, suggesting an ultimate origin by re-cycling of continental crust (S-type granites). These granitoid masses predate the major high-pressure metamorphism of the Pleasant Bay complex, as demonstrated by the occurrence of the characteristic assemblage oligoclase-hornblende-epidote in the contained dykes. The age of these old granite plutons is unknown, but they may perhaps be related to the dioritic plutons farther to the southeast, which have given a preliminary Rb-Sr date of 750 ± 65 Ma (Jamieson and Craw, 1983).

The Cheticamp pluton was emplaced in Late Precambrian or Early Paleozoic time (530 ± 44 Ma, Cormier, 1972, recalculated by Keppie and Smith, 1979). The chemical analyses of this pluton contain normative corundum, but the presence of primary hornblende, sphene and magnetite, and the broad range in SiO_2 content indicate a pluton developed by differentiation from an igneous protolith (I-type granite). The most basic observed phase of the Cheticamp pluton is diorite, but based on crosscutting relations and degree of deformation the dioritic to hornblendite plutons of unit Hm appear to be roughly coeval and possibly cogenetic. Cormier (1972) found that gneisses equivalent to the Pleasant Bay complex gave Rb-Sr ages slightly older than the Cheticamp pluton (550 Ma), which he interpreted to be metamorphic. This interpretation is entirely in accord with the present observations since the Cheticamp pluton clearly postdates the high-grade metamorphism of the Pleasant Bay complex, but precedes emplacement of the Jumping Brook complex.

The Cheticamp pluton contains acid and basic dykes some 100 Ma younger than the pluton (439 Ma) which appear to be feeders of the bimodal volcanic rocks of the Jumping Brook complex. The age of these dykes resembles that of the Cape Smokey pluton east of the Chéticamp area (453 ± 42 Ma, Cormier 1972, as recalculated by Keppie and Smith, 1979). These ages reflect an igneous episode of uncertain amplitude occurring between emplacement of the Cheticamp pluton and commencement of Silurian and Devonian activity. In the Chéticamp region this episode produced mafic rocks of tholeiitic affinity and acid rocks approaching minimum melts.

Silurian and Devonian igneous activity commenced with development of foliated and lineated granites cutting the Jumping Brook complex. These plutons characteristically contain muscovite and garnet, are extremely poor in dark minerals, and locally grade into their surrounding gneissic envelope. They appear in many cases to be anatectic melts which have moved only slightly from their site of generation. This impression is strengthened by the high initial Sr ratio of 0.709 reported by Cormier (1972) for the occurrence at the mouth of Mackenzie River.

These foliated to lineated bodies appear to have been emplaced roughly synchronous with metamorphism but the large plutons (Margaree and Salmon Pool plutons) are clearly post-tectonic, and appear very nearly contemporaneous to the Fisset Brook Formation. In this connection the following observations appear pertinent. (a) Debris of the granites occurs in the Horton Formation (Latest Devonian to Early Mississippian), but debris of the Fisset Brook Formation has not been recognized. (b) Dykes assigned to the Fisset Brook Formation by Neale (1963) and Jamieson and Craw (1983) cut the Margaree pluton (but may be cogenetic with it). (c) The radiometric ages of 370 and 376 Ma for the Fisset Brook Formation (Middle Devonian) clearly conflict with its paleontological age of Latest Devonian or Early Mississippian. The simplest resolution of these difficulties appears to be that the plutons are slightly older than the volcanics, the plutons having an age of perhaps 365 Ma, as suggested by Jamieson and Craw (1983) for the Margaree pluton, and the volcanics an age of closer to 350 Ma.

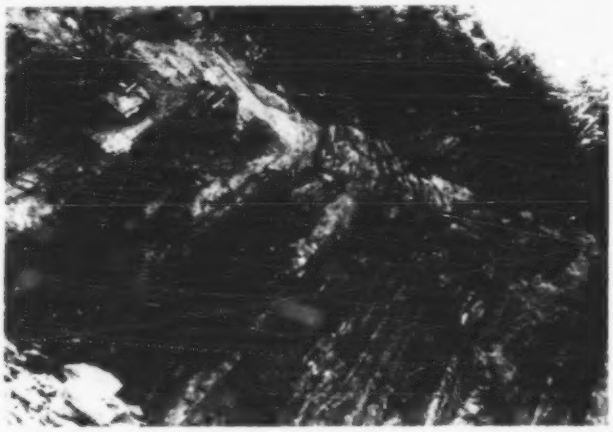


Figure 38a. Isocline in the Jumping Brook complex, upper part of Corney Brook. Height of exposure is about 8 m. (GSC 188581)

These considerations appear to define five periods of igneous activity. The oldest (Precambrian) formed the granite gneisses of the Pleasant Bay complex. Metamorphism associated with this episode (if any) has not been identified but boudined, metasomatized dykes suggest metamorphism before emplacement of the Cheticamp pluton. The second episode formed the Cheticamp pluton, and probably also small mafic plutons. This episode locally produced granulite facies assemblages in the Pleasant Bay complex, and erased all older radiometric ages in these gneisses, as shown by the occurrence of 550 Ma metamorphic ages reported by Cormier (1972). A third episode of igneous activity took place during the Ordovician (440-455 Ma), as shown by the age of the Jumping Brook complex volcanics and the Cape Smoky pluton. The extent of this episode is presently uncertain, but the striking similarity of the association leucocratic granite-bimodal volcanics to that of the later Devonian magmatism suggests that it could have been as extensive as the later Devonian magmatism. A fourth episode of magmatic activity produced the foliated and lineated garnetiferous leucogranite lenses (units **SDg** and **Dl**). This episode clearly was part and parcel of the metamorphism of the Jumping Brook complex. Finally a Middle to Late Devonian magmatism produced major leucocratic potassic granites, and subsequent bimodal volcanism. It may be remarked that there is a rather striking symmetry to the pairs: gneissic granite-amphibolite dykes; Cheticamp pluton-hornblendites; Cape Smoky pluton-bimodal Jumping Brook volcanics; and Devonian plutons-bimodal Fisset Brook volcanics.

In considering the nature of the granitic rocks, it may be remarked from Figures 31 and 32 that the great majority of the analyses fall within the region found by Winkler and von Platen (1961) to contain most analyses of granitoid rocks. Many of these analyses fall relatively close to the trace of the minimum melt composition for Q-Ab-Or. The exceptions are some of the Pleasant Bay granite gneisses, and salic volcanic rocks of the Fisset Brook Formation. For the latter, there is strong circumstantial evidence that metasomatism must be involved. The former is likewise assumed to reflect

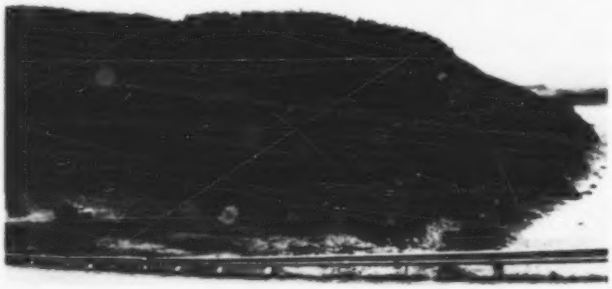


Figure 38b. Overturned isocline in Horton Group rocks south of Pillar Rock. (GSC 188560)

some metasomatism, in this case enriching the rocks in Na rather than K. It may be noted that in a general way the chemical analyses increase in K relative to Na with decreasing age. This would be compatible with melting at decreasing pressures. Figures 31 and 32 demonstrate that much of the chemical variation can be explained by fractionation or accumulation of a plagioclase of about An_{20} and potash feldspar. Such a process would produce the observed strongly linear arrays stretching between these two end-points. In the Cheticamp pluton, however, the variation seems better explained by fractionation of a very calcic plagioclase (near anorthite) and an alkali feldspar containing sub-equal amounts of sodic and potassic end-members.

STRUCTURAL GEOLOGY

Evidence of polyphase deformation

The gneisses and migmatites of unit Hpbp in the Pleasant Bay complex exhibit complex rootless isoclines a few tens of centimetres in length which lie in thin septa between schistosity planes (Fig. 4). Granite gneisses of the Pleasant Bay complex (unit **Hpbg**) truncate these migmatites, and show good foliation but little or no small scale folding. The Cheticamp pluton, which is brecciated but not gneissic, truncates the granite gneiss. First folding of the Pleasant Bay complex therefore predated emplacement of the Cheticamp pluton. The emplacement of the Cheticamp pluton marked a late to post-tectonic climax to strong metamorphism affecting the Pleasant Bay complex. Although two periods of development of foliation are clearly implied by small folded outcrop relations in the Pleasant Bay complex, in general, only one, pervasive, foliation could be mapped. This pervasive "main schistostity" of the Pleasant Bay complex has been termed S_m with the proviso that S_m is clearly composite. As noted S_m is locally axial planar to rootless isoclines (F_1).

The Jumping Brook complex unconformably overlies the Cheticamp pluton, and schistosity within the Jumping Brook complex (S_i) locally truncates an older schistosity (S_m, see Fig. 37). The Jumping Brook complex has been folded at least twice. Small, north-trending isoclines can be observed in many places (Fig. 38a), but the prominent folds on the map-scale are moderately tight north-trending upright folds which plunge north at moderate angles and are associated with strong cleavage. North of Chéticamp River, the outcrop pattern of Jumping Brook complex defines such a fold which refolds an older isoclinal fold whose nose can be observed along the Cabot Trail north of Grande Falaise. The older recumbent folds have been termed F_2, and the younger upright folds termed F_3 and their associated cleavage S_3. The isolated masses of mafic and metasedimentary schists within the Pleasant Bay complex appear to be cores of F_2 folds.

The older isoclinal F_2 folds cannot be demonstrated to affect the Pleasant Bay complex. They may represent a superficial style of deformation during which the Jumping Brook complex peeled off the underlying basement. This deformation apparently preceded metamorphism, since at the locality shown in Figure 38, the metamorphic minerals are unoriented (compare Fig. 11). The deformation producing F_3 clearly accompanied or outlasted metamorphism, since metamorphic minerals are found rotated into S_3, particularly in the higher grade parts of the Jumping Brook complex.

The Horton and Windsor Groups locally show tight to isoclinal folding, overturned in places (Fig. 38b). Such folds must be younger than F_3, since a strong mineral lineation associated with progressive metamorphism of the Jumping Brook complex, accompanies the latter, whereas the Horton and Windsor groups do not display metamorphism above sub-greenschist facies. These latest folds are here termed F_4.

The fold sequence presented here agrees essentially with the opinion of Wiebe (1972) who argued that on the east side of the Cape Breton Highlands the "George River Group?" (equivalent to the Pleasant Bay complex in the present terminology) had been folded prior to the deposition of the "Fourchu Group?" (equivalent to the Jumping Brook complex), from which it was separated by an unconformity. Similar opinions were expressed in the Cape North area by Neale and Kennedy (1975). Keppie (1979) generalized that throughout the Cape Breton Highlands, metasediments and metavolcanics unconformably overlie a pre-Hadrynian gneiss complex. MacDonald and Smith (1980) argued, contrary to these authors, that all the rocks of the northern highlands showed two, and only two, periods of deformation, and one period of kyanite grade metamorphism, thought to be Precambrian in age. These views may be correct for the Pleasant Bay equivalents, but successions exhibiting low pressure metamorphic series (andalusite-cordierite), and younger fold sequences (here termed F_3 and F_4) are clearly shown by radiometric dating and stratigraphic relations to be deformed and metamorphosed in Devonian or Silurian time, and these sequences appear to be of Paleozoic age.

Role of overthrusts and slides

In a previous paper overthrusts of post-Mississippian age were argued to play a role in the structure of northern Cape Breton Island (Currie, 1977). Within the mapped area particularly spectacular evidence of overthrusting can be seen at Grande Falaise where the thrust plane is exposed intermittently for 1 km just north of Jerome Brook, in Robert Brook where the Cheticamp pluton is thrust over shattered Fisset Brook Formation and Horton Group strata, and north of Grande Lac, where a hillock of Horton Group some hundreds of metres in diameter rests inverted on deformed Windsor Group strata. Together with other outcrops around the boundary of the Cape Breton Highlands, these localities suggest that thrusting or tectonic slides play a significant role in the present configuration of the geology, but they are insufficient to demonstrate the scale of motion or the causative mechanism.

A suggestion that the geology seen on the surface is insufficient to explain the geology at depth follows from the Bouguer anomaly gravity map (Earth Physics Branch Open File 79-7B). The Cape Breton Highlands as a whole exhibits a strong gravity high which locally reaches 300 milligals. A trough of low values follows the Margaree pluton. Since the

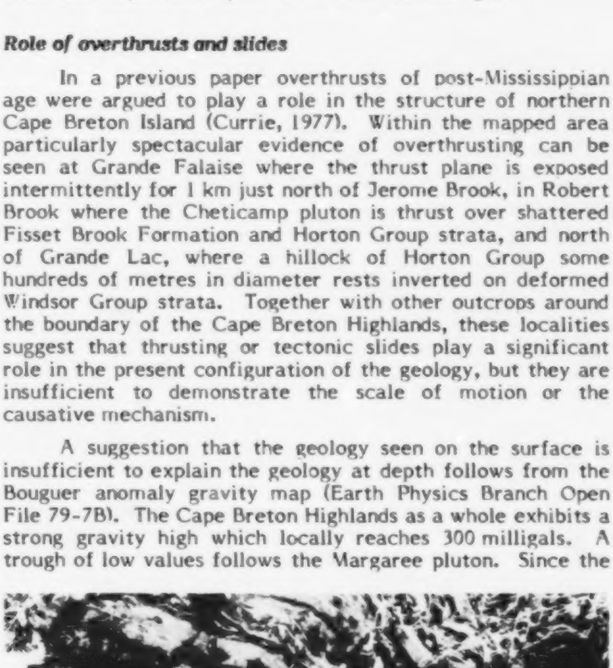


Figure 39. Banded mylonite northeast of Pillar Rock. Note the very well developed compositional banding, with small scale, rounded folds or eddies, and the sharp kink-folds deforming the banding. (GSC 188564)

exposed portion of the Cape Breton Highlands is almost entirely composed of quartzofeldspathic gneisses and granites of relatively low density, rocks of much higher density must be present at relatively shallow depths to explain the gravity anomalies. The size of the gravity anomaly seems to require a major mass of mafic to ultramafic rocks.

The magnetic anomaly map (Geological Survey of Canada, Map 7038G), accurately reflects the surface geology. The Jumping Brook complex and Margaree plutons cause magnetic highs, whereas the Pleasant Bay complex and Carboniferous rocks produce magnetic lows.

Development of cataclastic zones

The most spectacular feature of the structure of the Chéticamp region is the intricate network of steeply to vertically dipping cataclastic zones found along the sea-coast from Chéticamp River northward. These cataclastic zones cut both crystalline rocks and Carboniferous sedimentary rocks, but they are more prominently developed in the crystalline rocks. Zones of cataclasis vary from less than a metre to several hundred metres in width, and the rocks contained within the zones vary from phyllonitic material, through various types of breccias to mylonite, flinty crush rocks, and pseudotachylitic material. In a general way, the breccias occur in homogeneous coarse grained rocks, such as granites or amphibolitic rocks, whereas the banded phyllonitic and mylonitic rocks develop from rocks which had an initial planar structure. The cataclastic zones clearly are largest and most strongly developed along the western margin of the highlands, that is trending north-northeast. A zone of similar deformation with a similar trend follows the valleys of the Aspy and Margaree rivers, but along this zone many of the schistose rocks contain porphyroblastic chlorite, and some contain biotite, suggesting substantial recrystallization after deformation. Along the "Highland front" the banded rocks are uniformly fine grained, although many have a phyllonitic sheen due to myriads of minute, aligned chlorite and muscovite flakes. Between the two largest cataclastic zones occur a myriad of small, roughly parallel zones within the crystalline rocks. Such zones are generally 1 to 5 m in length and 10 to 30 cm wide. The rocks within the zones are markedly finer grained and more foliated than those around them, although the mineralogy is generally similar. In some cases the fine grained schistose zones may be rich in chlorite and hematite relative to the surroundings. Such zones by no means always have north-northeasterly trends. In fact the most dominant trend is a north-northwesterly trend, very well exemplified in the branches of Fishing Cove River. In some cases these zones clearly represent faults, for example the valley extending north-northwest from Chéticamp River to Pillar Rock, along which units as young as the Horton Group are clearly offset. This valley also exhibits the characteristic left-hand curvature of the ends of these zones, which typically gives them a sigmoid shape. The two complementary sets of cataclastic zones, one trending north-northeast and the other north-northwest, dominate the structure of northwestern Cape Breton Island.

The history of development of these zones is not clear. The Margaree pluton clearly truncates the cataclastic zone following the Aspy River, although the pluton is itself slightly deformed along the margins. The development of the zones therefore commenced before, or during, emplacement of the late plutons. This emplacement postdated the major metamorphism of the crystalline rocks, but according to Brun and Cobbold (1980) shear heating in major cataclastic zones could readily exceed 100°, which together with heat of crystallization from the pluton would be quite adquate to produce the observed recrystallization. Along the western front of the highlands, the cataclastic zones are known to deform the Windsor Group (as evidenced by numerous veins, lenses and inclusions of gypsum and carbonates in the cataclastic zones), and may postdate the youngest sedimentary rocks, since Cameron (1948) reported numerous faults cutting these rocks. Despite their relative youth, the cataclastic zones exhibit small-scale cross structures, and local recrystallization. The small kinks and internal folds deforming the mylonitic banding (Fig. 39) may reflect fluid-like conditions in the fine grained paste, while the observed recrystallization could be ascribed to shear heating (Brun and Cobbold, 1980).

The significance of the cataclastic zones for structural reconstruction of western Cape Breton Island must be assessed in light of other evidence. The front of the highlands clearly marks a major boundary in the Carboniferous section. Up to 5 km of Carboniferous section accumulated in the Gulf of Saint Lawrence (Howie and Barss, 1975), whereas contemporary conglomerates around the edge of the highland show that it was being sub-aerially eroded at this time. Hence there appears to be a vertical displacement of more than 5 km between the highlands and the Gulf of Saint Lawrence, some or all of which probably was localized in the cataclastic zones. It may be remarked that this displacement is in the proper sense to bring up progressively deeper rocks toward the centre of the highlands, as suggested by the studies of metamorphic grade and the positive gravity anomaly.

It cannot be assumed that movement on the cataclastic zones was entirely, or even predominantly vertical. Where small scale displacements can be observed across cataclastic zones, they are consistently in the left-hand sense. Such small-scale displacements are particularly common across the north-northwest trending zones. The highland front marks the northwest edge of the crystalline rocks. All shoals offshore consist of Carboniferous sedimentary rocks. It could therefore be hypothesized that the complex of shear zones represents a major zone of transcurrent left-hand movement, separating two distinct geological terranes. On this view, the vertical movements during the Carboniferous, with their resultant complex pattern of sedimentation, represent relatively small scale side effects of major transcurrent motion. The association of pseudotachylite with the cataclastic zones likewise suggests transcurrent motions (Philpotts, 1964).

If the concept of major transcurrent motion is the correct one to apply in this case, close relation of transcurrent motion to other geological events is required. Since commencement of the cataclastic zones predated emplacement of the Margaree and Salmon Pool plutons, a moving plate may have carried along, or generated, the relevant magma. The same considerations apply to the Fisset Brook Formation. The elongated, yet unfoliated character of these bodies could then be assigned to emplacement along cataclastic zones in a state of tension. A glance at the map suggests that this concept is quite reasonable. The Margaree pluton occupies a position parallel to the dominant north-northeast trending zones, while the Salmon Pool pluton and Fisset Brook Formation occupy zones parallel to the subsidiary north-northwest trending zones.

Comparing earlier thermal events to deformation, note that within the Jumping Brook complex porphyroblasts overprint an early schistosity in low grade rocks, but are themselves overprinted and rotated in higher-grade rocks. These relations suggest that the sedimentary pile heated passively until metamorphism commenced, and that as heating

Table 6. Geological development of the Chéticamp region

Date (Ma)	Period	Deposition	Plutonic events	Deformation	Interpretation
	Carboniferous	Riversdale & Canso groups			
				mainly transcurrent faulting, cataclasis	plate motions
		Windsor Group			
				uplift, local, thrusting	
		Horton Group			molasse
350	Devonian	Fisset Brook Fm (IB_3)		rifting, uplift	thermal readjustment relaxation of compression
			Margaree and Salmon Pool plutons (IG_3)		anatexis in thickened pile
			metamorphism (M_3) and anatexis	tight north plunging folds (F_3)	compression of sedimentary wedge
	Silurian			isoclinal recumbent folds (F_2)	"thin-skinned" tectonics
440	Ordovician	Jumping Brook complex	mafic dykes (IB_2) Cape Smoky pluton		break up of craton subsidence, melting of deep crust continental collision(?)
530	Cambrian Hadrynian		Cheticamp pluton (IG_2) metamorphism M_2 and anatexis(?)	completion of composite fabric (S_m)	orogeny continental collision?
	?Hadrynian		mafic dykes (IB_1) Pleasant Bay granites (IG_1) metamorphism (M_1?) mafic dykes (IB_0)	commencement of composite fabric S_1	
	?Helikian	Pleasant Bay complex			volcano-sedimentary cover on subsiding craton
			Anorthosite		basement(?)

M - period of metamorphism, numbered from oldest (1) to youngest (3)
I - period of igneous activity, IB for basic magmatism, IG for salic magmatism numbered from oldest (1) to youngest (3)
F - period of folding numbered from oldest (1) to youngest.

intensified the rocks became more mobile. Note that some early granites (units **SDg** and **Dl**) were emplaced under a notable stress field. On the transcurrent motion model, it would seem reasonable to suppose that emplacement of these bodies records one of the last events before the terrane decoupled itself, and stress within the block fell to very low values because strain was concentrated in cataclastic zones along its edges.

Major mylonite zones have also been reported which do not involve large vertical or transcurrent motions, but rather are related to major folding (Carreras et al., 1980). The mechanism advocated by Carreras et al. (1980) does not seem applicable to the Cape Breton Highlands because (a) the cataclastic zones appear to cut indifferently across folds, and do not change dips on opposite sides of fold axes, and (b) at least some of the granites clearly postdate the activation of the cataclastic zones, and (c) many of the cataclastic rocks display open voids, vugs, and other evidence of formation at shallow depths, suggesting that development of the cataclastic zones, although perhaps commencing at depth, persisted virtually to the surface.

On balance, it seems probable that the cataclastic zones represent loci of major movements, probably horizontal, almost certainly vertical. Comparison with the similar Belleisle fault in New Brunswick (Currie, 1984) and Cabot fault in Newfoundland suggests that a continuous belt of cataclastic zones may form a major geological boundary in the Canadian Appalachians.

Structural synthesis

Table 6 gives a summary of the geological development of the Chéticamp region. A series of interpreted cross-sectional cartoons is included in Figure 2. Due to the large scale remobilization during the several episodes of metamorphism, and the pervasive effect of late cataclastic deformation, the early parts of the structural development are not known with precision. However the oldest small structures in the region must be the rootless isoclines in migmatites of the Pleasant Bay complex (Fig. 5). The granitic gneisses of the Pleasant Bay complex clearly truncate these structures, and hence must be younger. Since these granitic bodies themselves exhibit a strong foliation, the fragmentary evidence suggests two periods of deformation prior to the emplacement of the (unfoliated) Cheticamp pluton. This assessment leaves aside the presence of basement to the Pleasant Bay complex.

Consideration of the metamorphic grade of the Pleasant Bay complex suggests that emplacement of the Cheticamp pluton marked the end of a period of major deformation and metamorphism which welded together and essentially homogenized all older rocks, leaving the prominent composite foliation S_m. The antique gneisses of the Pleasant Bay complex retain within them enclaves of pelitic and carbonate rocks, as well as sodic gneisses suggestive of intermediate volcanic rocks. The basement upon which these supracrustal rocks accumulated cannot now be definitely identified, but the anorthosite sliver of the Red River complex, now tectonically enclosed in the surrounding gneisses, does not form part of the supracrustal sequence, and may be a fragment of basement.

The cycle of deposition, metamorphism and plutonism culminating in emplacement of the Cheticamp pluton was followed, after an interval of about 100 Ma, by development of a volcano-sedimentary complex upon it (Jumping Brook complex), and farther to the east by the emplacement of a major leucocratic granitoid pluton (Cape Smoky pluton). Between the emplacement of the Cheticamp pluton, at conditions deduced to be close to 650°C, 900 MPa, (9 kilobars), the pluton and surrounding Pleasant Bay complex must have undergone major uplift and erosion to bring them to the surface where the Jumping Brook complex could be laid down upon them. The zones along which differential movement took place cannot be confidently identified, but the younger cataclastic zones so prominent along the western edge of the highlands commonly developed in strongly schistose augen gneisses, suggesting that the younger movement zones may have re-activated older, more or less recrystallized movement zones. The bimodal character of the Jumping Brook volcanism is compatible with an origin by rifting of a craton.

Subsequent to the deposition of the Jumping Brook complex these rocks were deformed twice (F_2, F_3). The earlier deformation predated metamorphism of the Jumping Brook complex, produced small, west verging isoclines (Fig. 38), did not affect the older Pleasant Bay complex, and produced no recognizable cleavage. These relations suggest that this deformation may have been of a superficial, thin-skinned character. F_2 folds are clearly overprinted by porphyroblasts in the Jumping Brook complex (Fig. 11), which in turn are rotated into the strong cleavage associated with F_3 (Fig. 12). The exact dates of these two deformations cannot be fixed precisely, but if the age of the Jumping Brook complex is accepted as late Ordovician or early Silurian, and the age of the foliated granite at the mouth of Mackenzie River is taken as middle Devonian, as suggested by the radiometric dates, then both deformations must be of Silurian to early Devonian age. Analysis of the metamorphic grade shows that emplacement of the foliated granites occurred near the peak of metamorphism. The intensity of the deformation increased with metamorphic grade, as shown by the varying relations of porphyroblasts to cleavage (Fig. 11, 12).

The major plutons (Margaree and Salmon Pool) clearly postdate most or all of the major deformation and metamorphism. The apparent synchroneity of the major plutons and basaltic volcanism of the Fisset Brook Formation suggests that this time was one of tension, rather than compression. Analysis of the cataclastic zones raised the possibility that this radical change in structural regime represents transition to translational motion of a large crustal block. The previous section detailed how this regime could produce rifting and overthrusting in adjacent areas, and helps to explain the observed complexities of Carboniferous sedimentation. The observed pattern of cataclastic zones (very strong and large north-northeast trending zones, subsidiary north-northwest trending zones exhibiting left-hand displacement) is compatible with motion of a Cape Breton "block" relatively northward due to oblique compression from the south-southeast. The tectonic pattern observed during the Silurian to Carboniferous bears a remarkable resemblance to that exhibited by the Pleasant Bay complex and Cheticamp pluton. In both cases a volcano-sedimentary complex accumulated upon rather ill-defined contenental basement. In both cases deformation and metamorphism culminated in emplacement of syntectonic (foliated) granites. In both cases emplacement of the foliated granites was followed after a very short interval by emplacement of unfoliated granites. These analogies suggest analogous causes.

TECTONIC SETTING OF THE CHÉTICAMP REGION

Pre-Carboniferous tectonic setting

Since the pioneering paper of Bird and Dewey (1970) the structure of the Appalachian orogen has been interpreted in terms of plate tectonic models. According to current models of the northern Appalachians (Williams, 1979), development of this region can be divided into an initial period of continental rifting and opening of the Iapetus ocean sometime in the late Precambrian, followed by plate consumption which closed the Iapetus Ocean, probably in mid-Ordovician time. The Siluro-Devonian Acadian orogeny is presently thought by most Canadian authors to postdate ocean closure. In their simplest form, models of the opening and closing of Iapetus do not explain latest Precambrian to Cambrian metamorphism and granite emplacement, but several authors (Rast, 1979; Keppie, 1981; Hada, 1980) have attributed latest Precambrian metamorphism and granites to subduction of material beneath the proto-European or proto-African craton, a concept which can be made compatible with the opening and closing of Iapetus. Post-Devonian deformation of the northern Appalachians cannot be easily fitted into the Wilson cycle, and is here treated as a separate problem.

With the exception of very broad scale syntheses such as that of Williams (1979), no effort has previously been made to fit northwestern Cape Breton Island into models of the northern Appalachian orogen. Broad scale syntheses of the Canadian Appalachians (Williams, 1979) recognize four major belts of rocks outward from the North American craton, namely the collapsed Hadrynian to early Paleozoic sedimentary accretion prism lying on the thinned edge of the craton (Humber zone), relicts of oceanic material and superincumbent volcanic arcs (Dunnage zone), an intensely deformed and metamorphosed accretionary wedge possibly related to an eastern continent (Gander zone), and a generally supra-crustal volcano-sedimentary sequence of Hadrynian to Ordovician rocks, locally known to rest on continental basement (Avalon zone). For the present purpose the farthest outboard zone of lower Paleozoic clastic sedimentary rocks (Meguma zone) is not relevant. These zones can be most unequivocally recognized in Newfoundland, but attempts have been made to extend them the length of the orogen. The rocks of the Cape Breton Highlands, high-grade gneiss and schist of several ages intruded by megacrystic granite, most closely resemble the Gander zone. However certain difficulties with this identification have led to proposals that the highlands may be (an anomalous) part of the Avalon zone. As will be shown, these two proposals are not mutually exclusive. Objections to assigning the rocks to the Gander zone are four-fold. First, the highlands contain granitoid rocks of Cambrian or late Precambrian age. Such granites have hitherto been observed only in the Avalon zone, and are thought to be characteristic of terranes not part of the American craton during the life of the Iapetus ocean. Second, the highlands clearly contain Precambrian gneisses, as well as anorthosite typical of the American craton. No Precambrian basement to the Gander zone of Newfoundland has yet been confirmed, although isotopic studies of younger granites suggest that material as old as 1100 Ma may be present at depth (Currie, 1983). Third, the Cape Breton Highlands exhibit a fringe of Carboniferous rocks. No Carboniferous supracrustal rocks have been recognized in the Gander zone of Newfoundland. Finally, and perhaps most significant, the continuation of the Gander zone of Newfoundland has generally been identified with rocks far to the west of Cape Breton Island in central New Brunswick and Maine. Resolution of these difficulties requires significant adjustment of current large scale models of the Canadian Appalachians. Some possible, speculative solutions for the objections noted above are discussed in the following paragraphs.

The age of opening of the Iapetus ocean remains uncertain. Isotopic evidence from the Cape Breton Highlands (Jamieson and Craw, 1983) and the Saint John region of New Brunswick (Olszewski et al., 1980) suggests a major igneous episode at about 800 Ma which emplaced ultramafic to intermediate plutons. Studies of dykes supposedly associated with continental break-up suggest similar ages (Strong and Williams, 1972; Rast, 1979). The dominantly dioritic material of the igneous activity might be explained by strong upwelling of hot mafic material into the crust (plume-type activity) producing anatexis and magma mixing. I assume that the region broken up was a passive continental margin veneered by stable-shelf sediments, as demonstrated by the succession around Saint John, New Brunswick (Currie et al., 1980), and that during break-up volcanic material, predominantly of dioritic (that is andesitic) composition was laid down on this succession as a result of the widespread igneous activity. The sedimentary parts of the Pleasant Bay complex and the presumably correlative George River and Green Head groups represent the sedimentary succession, while volcanic rocks of the Pleasant Bay complex testify to volcanism during continental break-up.

Such a catastrophic break-up would probably not produce a single smooth boundary. The contemporary examples of the Horn of African and Baja California suggest that fragments tens or hundreds of kilometres in dimensions would be created, which would be carried away from cratonic America, gradually subsiding due to cooling, thin continental crust, and possibly emplacement of mafic igneous complexes during break-up. Possible initial sub-aerial clastic sedimentation would thus be succeeded by marine sedimentation. King (1979) and Rast (1979) explained the Avalon zone by continental fragmentation, and subsequent sedimentation around the fragments, in agreement with the present proposals, but they did not propose an American origin for the fragments. Many years previously, in the Saint John region, Alcock (1938) correlated Precambrian mioclinal sediments underlying supracrustal rocks of the Avalon zone with the Grenville province of the American craton, but more recent writers have tended to assume a non-American provenance for all parts of the Avalon zone. However the massif-type anorthosite of the Cape Breton Highlands seems to offer an argument for deriving at least this block from the American craton, since the only other massif-type anorthosites around the North Atlantic basin, those of Scandanavia, are also thought to be part of a fragment split off from America (Wynne-Edwards and Hasan, 1968). The advantages of assuming an American provenance for the Avalon zone can be seen by considering the fate of the assemblage of continental fragments and their various veneers.

Large scale deformation, metamorphism and plutonism occurred at 520-600 Ma in Norway, western England and France, and emplacement of granites in the Avalonian terrane has commonly been correlated to these events (Rast, 1979; Keppie, 1981). Stratigraphic evidence from western Newfoundland, however, shows that by Arenig time (485 Ma) fragments of ocean floor were being transported westward onto the American craton (Stevens, 1970). The time span between emplacement of the Cheticamp pluton (550 Ma) and mid-Ordovician closure of Iapetus seems very short to allow for a return trip of the Cape Breton Highlands across the Iapetus Ocean, or for two complete Wilson cycles if the Avalon zone is assumed to derive from some source other than America. A simpler mechanism would have the Avalonian assemblage of fragments stop some relatively short distance offshore of America, in the manner of Japan off Asia. This would allow Eocambrian-Cambrian sediments of stable shelf type to accumulate inboard of the Avalonian assemblage, while the Avalonian terrane lay above a subduction zone. If Avalonia were underlain by a west-dipping subduction zone, the westward increase in metamorphic grade across the Avalon zone into the Gander zone finds a natural explanation. Note that the trace of this zone would have to lie east of the present continental margin. If an eastward dipping subduction zone is assumed, this westward increase in metamorphic grade, which appears to be characteristic along the whole length of the Avalon zone, is difficult to explain. The possible explanation by major transcurrent faulting between the Avalon and Gander zones (Hanmer, 1981) does not explain the presence of a 530 Ma granite in highly metamorphosed rocks of the Cape Breton Highlands. Jenness (1963) argued for direct correlation of rocks of the Gander and Avalon zones in Newfoundland, and similar views have been expressed in Cape Breton Island (Milligan, 1970) and New Brunswick (Currie et al., 1980).

The Iapetus ocean was essentially closed by Caradoc or Ashgill time, with the exception of narrow sea-ways or embayments (Williams, 1979; Currie et al., 1980). Significant metamorphism and plutonism did not commence until Middle or Late Silurian time, a lag which can be ascribed to the heating time for a thickened sedimentary pile (Currie and Pajari, 1981; Pajari et al., 1981). During this time the Jumping Brook complex accumulated on the Cape Breton Highlands. Note that the highlands had been metamorphosed at high temperatures and pressures about 550 Ma (Cormier, 1972), and hence must have been at substantial depth, and subsequently had risen and been eroded prior to deposition of the Jumping Brook complex, removing any remnants of "Avalonian cover" which may have existed. Hence if the highlands is considered as part of an Avalonian terrane, it must be basement to the Avalonian supracrustal sequences observed in Newfoundland and southeastern Cape Breton Island. The scale and rapidity of the vertical movements strongly suggest plate-convergent processes. The mechanical reasons for renewed subsidence, permitting accumulation of the Jumping Brook complex cannot be explained at present, but an exactly analagous and contemporary sequence of ocean closure, followed by bimodal vocanism and accumulation of marine sediments is well documented in the Indian Islands Group of northeastern Newfoundland (Currie et al., 1980). Perhaps relaxation of compression, or tensional re-adjustments perpendicular to compression, may be responsible.

Effects of self-heating in the thickened sedimentary pile following the closure of Iapetus have been examined by Currie and Pajari (1981) and Pajari et al. (1981). In addition to the anatexis and metamorphism considered there, self-heating also has important kinematic effects. Brun and Cobbold (1980) suggested that a localized rise in temperature of 100° is sufficient to constrain subsequent deformation to the heated zone. I assume that self-heating was relatively homogeneous up to at least chloritoid porphyroblasts. This does not imply completely homogeneous stress, since the presence of two ages of folding in the Jumping Brook complex strongly implies directed stress, but rather that the strength of the rocks was sufficient to support the stress until heated above a critical temperature. At this point I assume loss of strength, and rapid concentration of strain into narrow zones as noted by Brun and Cobbold (1980), giving rise to the cataclastic zones so typical of the Cape Breton Highlands. As previously noted, the orientation of these zones, and relative motion, suggests oblique compression from the southeast, which would be compatible with oblique subduction of material beneath the Avalon zone prior to self heating. Hence the development of the cataclastic zones does not necessarily imply a radically new style of deformation, but rather evolutionary change of the previous style.

Carboniferous and later tectonic setting

The presence of Carboniferous and younger deformation in the Appalachians has been recognized for many years. Poole et al. (1970) drew attention to the concentration of this deformation in a narrow zone. Some parts of this zone have long been suspected to be associated with major transcurrent movements (Wilson, 1966; Currie, 1984), but at least three differing types of explanation have been offered for this deformation zone. These may be termed (a) the normal movement model (Poole et al., 1970), (b) the convergent movement model (Rast and Grant, 1973), and (c) the transcurrent movement model (Keppie, 1981). The relevant features of these models, as applied to the Cape Breton Highlands, are reviewed below.

The normal movement model assumes essentially vertical Carboniferous motions affecting generally autochthonous rocks ("Maritime disturbance" of Poole et al., 1970). On this model the whole of the Cape Breton Highlands could represent an uplifted block. The gravity high suggests thin crust, due either to erosion from uplifted blocks or to rifting and intrusion. The observed overthrusts could be ascribed to minor tectonic slides at the edges of uplifted blocks, presumably lubricated by gypsum beds. The direction of transport of thrust blocks agrees with this concept, westward at Grande Falaise and Grand Lac, southward in the Margaree valley, and southeast at Saint Anns (Currie, 1977). The Fisset Brook Formation could be considered to erupt from rifts. The major objections to the model are (a) local evidence for intense post-Devonian thrusting accompanied by metamorphism in southern New Brunswick (Rast and Grant, 1973), and (b) the very narrow elongate zone of deformation (Currie, 1983). The "Maritime disturbance" cannot be compared to a basin and range situation because of this elongate character, nor to graben tectonics because of the local presence of thrusting and metamorphism. The model thus appears mainly descriptive, with little explanatory power.

An alternative solution was proposed by Rast and Grant (1973), based on observation of large scale Carboniferous thrusting in southern New Brunswick. They assumed the zone of thrusting to represent the edge of a major sheet of allochthonous rocks derived far to the east, and proposed that these metamorphosed rocks were related to the Permo-Carboniferous Variscan orogeny of Europe. The model assumed that such thrust and metamorphosed Carboniferous rocks represented outliers of the products of a major period of compressive deformation, metamorphism and plutonism, presumably related to convergence of two plates. This model now appears inadequate to explain the observations in southern New Brunswick (Currie, 1984), and has little appeal in northern Cape Breton Island. Undoubtedly correlative Carboniferous rocks of the Horton and Windsor groups outcrop at intervals around the entire circumference of the Cape Breton Highlands. Hence if the crystalline rocks are a thrust, allochthonous sheet, the sheet cannot be thicker than the height of the highlands, a maximum of 450 m. Such a thickness is too little to explain the negative gravity anomaly over the Margaree pluton. The model likewise fails to account for the observed positive gravity anomaly, since it predicts a thickened, rather than thinned, continental crust. Finally, the lack of tectonic fabric in the Carboniferous sedimentary rocks of the Cape Breton Highlands suggests superficial, rather than deep-seated deformation.

A third type of model has recently come into favour due to paleomagnetic observations favouring large paleolatitude differences between parts of the Canadian Appalachians (Kent and Opdyke, 1980; van de Voo et al., 1980). In its extreme form the model assumes the Canadian Appalachians to be formed of a large number of disparate terranes separated by transcurrent faults of major displacement (Keppie, 1981). This version suffers from major difficulties in placing the required faults so as to agree with geological observations. In a more modest form, assuming only two or three major fragments, and a combination of transcurrent and convergent motions ('transpression' and 'transtension' of Reading, 1980), this type of model offers an attractive and observationally supported model (Currie, 1983, 1984). The model can explain both uplift and subsidence in a small region (Reading, 1980), and hence offers a reasonable explanation for the complex pattern of Carboniferous sedimentation. Small scale rifting and volcanism (Fisset Brook Formation) and nearby, virtually contemporaneous thrusting can be explained in the same fashion. The major objection to such a model for the Cape Breton Highlands is the weak evidence for a large transcurrent fault along the western edge of the island. However, as previously noted, the Cape Breton Highlands lies in a narrow, elongate zone along which evidence for transcurrent motion is locally very strong. An interesting peculiarity of this zone is the unusual relation to Devonian granitoid and rhyolitic rocks. In New Brunswick the Belleisle fault separates a terrane to the northwest containing Devonian granitoids from one to the southeast virtually lacking such rocks (Currie, 1984). The Cabot fault of Newfoundland separates a terrane to the northwest which contains Devonian granitoid rocks from one to the southeast in which only Silurian granitic rocks have been recognized, whereas the Cobequid fault of Nova Scotia separates terranes which both contain Devonian granitoid rocks.

The presence of such diverse terranes suggests that the net effect of post-Devonian movement has been to transpose diverse terranes, and shuffle together fragments originally separated by tens or hundreds of kilometres (compare Tipper et al., 1981). The zone of deformation outlined above appears to cut obliquely across the trend of the Avalon and Gander zones, such that the Gander zone is essentially completely cut out south of Cape Breton Island, and the Avalon zone is cut out in southern New Brunswick. Equivalents of these zones cannot be certainly identified to the northwest of the faulted zones, and it would seem more prudent to regard possible correlatives as separate terranes with possibly differing tectonic, metamorphic and intrusive histories.

Summary

The classical idea of an opening and closing of the Iapetus Ocean as a generator for Appalachian geology can be applied to the Chéticamp region, but is not adequate by itself to explain observations. The following speculative notions appear to have some value in explaining the local and regional geology;

1. The Avalon zone consists of fragments of the edge of the American craton variously veneered by a Hadrynian or Helikian mioclinal assemblage, and a younger volcanic and sedimentary assemblage associated with opening of the Iapetus Ocean.
2. The Avalon zone is not particularly far travelled, but in Cambrian time lay off the coast of America, separated from it by a marginal sea.
3. Metamorphism of the Avalon rocks and emplacement of granite resulted from relative motion of proto-Europe towards America, with plate consumption on a west-dipping subduction zone lying seaward of Avalon zone.
4. The mid-Ordovician ocean closing recorded in Newfoundland refers to the closing of the marginal sea behind the Avalon zone. Other parts of the orogen may record closing of various small sea-ways at slightly varying times and with differing histories of deformation and metamorphism.
5. During self-heating following ocean closure, roughly the same stress field persisted as had effected ocean closure. When the rocks were sufficiently intensely heated zones of high strain developed due to interaction of thermal and mechanical effects.
6. The net effect of these high strain zones has been to displace terranes relative to one another by substantial amounts (tens or hundreds of kilometres), but not to alter significantly the broad scale zonal structure of the orogen.

ECONOMIC CONSIDERATIONS

Summary of known mineral occurrences

A variety of mineral occurrences of economic interest has been reported within the mapped area. Most of these occurrences have been known for many years, and little work has been done on any of them in the past 10 years. The following comments have been compiled for the most part from assessment files of the Nova Scotia Department of Mines and Energy, supplemented by field observations.

The oldest reported workings in the Chéticamp region are termed the Mine à Pascal, a few hundred metres southeast of the hamlet of Plateau. Two shafts were driven here before 1865, and copper is said to have been extracted (presumably from chalcopyrite). The locality lies very close to a thrust contact of acid and basic volcanics of the Fisset Brook Formation over sandstone of the Horton Group. A small amount of green copper stain can be seen in red siltstone intercalations in the Fisset Brook Formation. Shallow drilling by Phelps Dodge in 1966 at this site intersected very low grade, probably supergene, covellite-chalcopyrite mineralizaton in red siltstone interbedded with rhyolite. The highest recorded assay was 0.2% Cu over 5 feet (1.5 m).

The best known and most thoroughly investigated group of prospects lies in or near Faribault Brook and nearby in Daphinée Brook. The main showing in Faribault Brook (Galena Mine) was discovered in 1897, explored by adits in 1898 and 1901, further explored in 1927, and again intermittently explored from 1950 to 1964. The rocks are interbedded chloritic quartzites, phyllites and acid volcanics of the Jumping Brook complex which have been strongly sheared, and locally silicified and epidotized. Mineralization occurs as lenticular masses lying in the foliation plane, and strongly streaked out on the lineation, and as younger gash veins up to 2 m in length which cut the foliation at high angles. The older mineralization tends to be rich in pyrite and pyrrhotite and low in Cu, Pb, Zn and Ag. The mineralization clearly follows certain compositionally favourable horizons. The best assays from the Galena Mine are reported to be about 2% Pb, 0.14 oz/ton Au and 26.6 oz/ton Ag over 5 feet (1.5 m).

Other properties farther up Faribault Brook are similar in character (Iron Cap showing, Mountain Top), although one zone containing slightly more Cu has been reported (Grandin Zone). None of these showings are considered to be of current economic significance.

A similar but slightly higher grade showing was discovered in 1953 in Daphinée Brook, a few hundred metres east of the Faribault Brook showings. The mineralization of sphalerite + galena + pyrite + chalcopyrite occurs both massive and disseminated, apparently associated with a garnetiferous bed in the Jumping Brook complex, but also associated with strong north-trending shears. A chip sample across a width of 10 feet (3 m) returned 0.038 oz/ton Au, 3.8 oz/ton Ag, 1.73% Pb, 4.8% Zn and 0.06% Cu. The mineralization is commonly associated with large amounts of carbonate as gangue. A similar occurrence was seen during the present investigation in the upper part of Corney Brook. Older accounts consistently mention arsenopyrite, but this mineral was not seen during the present work.

A different type of base metal mineralization occurs on Rocky and Campbell brooks a few kilometres south of Faribault Brook. The Rocky Brook occurrence lies in brecciated and altered basalt and rhyolite of the Jumping Brook complex, in which silicification, epidotization and chloritization are widespread. The rocks are cut by porphyry dykes probably correlated with the Margaree granite. The mineralization follows certain compositionally favourable beds, with disseminated pyrrhotite, chalcopyrite, galena and sphalerite. Later gash veins appear to have remobilized part of this material resulting in partial replacement of pyrrhotite by pyrite. Extensive drilling on this occurrence gave disappointing results with the best assays giving about 1.5% Cu and 0.7% Zn over 5 feet (1.5 m). The Campbell Brook occurrence, which lies near the headwaters of the brook, resembles the Rocky Brook occurrence but contains negligible amounts of Cu.

Mapping during this study discovered additional minor Cu mineralization of the Mine à Pascal type in the Fisset Brook Formation in Robert and Jerome brooks, and in Horton Group rocks at Pillar Rock.

Gypsum was mined for many years from a series of large quarries in the Windsor Group about 3 km east of Chéticamp. These quarries lie in intensely folded rocks right at the base of the escarpment fronting the highland. The quarries closed when the anhydrite to gypsum ratio became unacceptably high. Gypsum and gypsum-calcite veining, possibly derived from remobilized Windsor Group rocks, also occurs in Robert Brook and at Grande Falaise. An unusual, possibly related type of mineralization occurs from Grande Falaise north to Corney Brook near the Cabot Trail. Barite is disseminated in brecciated Corney Brook and Jumping Brook rocks, in places associated with green fluorite and small amounts of calcite and galena. Just south of Corney Brook this material has been quarried from a small pit east of the Cabot Trail. Large amounts of anhydrite and gypsum also occur north of Pleasant Bay along the seacoast, and gypsiferous carbonate lenses occur in the Pleasant Bay complex along Grande Anse River.

Relation to structural features

The occurrences of mineralization in the Chéticamp region share a number of significant similarities. All occurrences lie either in a narrow north-northeast trending zone, roughly coincident with the front of the Cape Breton Highlands or in a parallel zone along the Aspy and Margaree rivers. This generalization applies both to occurrences in the Jumping Brook complex and the Mississippian rocks. The base metal occurrences are confined to intercalated volcanic rocks and metasedimentary schists, whether of the Jumping Brook complex or Fisset Brook Formation. All mineralization, whether metallic or non-metallic, has been remobilized during later deformation. All mineralization is associated with significant amounts of carbonate, although carbonate is otherwise quite rare, particularly in pre-Carboniferous strata.

The association of base metal mineralization with intercalated volcanic and metasedimentary schists suggests a possible syngenetic origin. The mineralized zones appear to contain an unusually high proportion of acid volcanic material, but the more mafic parts of the host rocks contain Cu mineralization, whereas the more acid host rocks contain Ag-Pb mineralization. The mineralized zones lie close to, or perhaps within, major cataclastic zones which bound and divide the Cape Breton Highlands. The mineralized zones in Faribault Brook and northward may also have been affected by late tectonic slides involving Carboniferous rocks. The mineralized zones along Faribault and Daphinée brooks appear to be truncated to the north by a cross-fault along the Chéticamp River, and the mineralization in Campbell and Rocky brooks may also be truncated to the north by the cross-fault which offsets the Margaree pluton in this region.

General considerations suggest that significant base metal discoveries are not likely in the Chéticamp region. Mineralization occurs mainly in the low grade parts of the Jumping Brook complex, and adjacent parts of the Fisset Brook Formation. No mineralization has been observed in rocks of staurolite or higher grade. The occurrences so far explored all appear to have been remobilized during shearing and faulting. However the low grade parts of the Jumping Brook complex are relatively accessible, and the zones of deformation can be readily identified by their topographic expression. These zones have already been fairly well explored. It thus seems unlikely that major new discoveries will be made in this region, although correlatives of the Jumping Brook complex elsewhere in the highlands may well be worth exploring.

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Appendix 1. Location and petrography of specimens from the Chéticamp region

Specimen	Easting (NTS 11K)	Northing	quartz	microcline	perthite	orthoclase	plagioclase	An(mol)	biotite	muscovite	hornblende	actinolite	garnet	epidote	chlorite	staurolite	chloritoid	cordierite	Al-silicate	sphene	apatite	carbonate	zircon	tourmaline	magmetitic	diopside	pyrite	scapolite	Description
Biotite gneiss and schist (unit Hpbp)																													
3-1	696610	5191350	x				x	32	x					y							z		z		z				Mesocratic, homogeneous foliated biotite gneiss
21-2	658260	5176450	x				5	28	x		y		y	y	y						z			z					Weakly foliated mesocratic biotite gneiss
45-1	665900	5179380	y				x	37	x		x			x						y	z								Mesocratic biotite gneiss with hornblende seams
50-1	662190	5184590	x			y	x	34	x				y								z			z					Leucocratic quartz-feldspar biotite gneiss
53-2	664250	5184630	x			y	x	30	x		y		y	y	y					y									Homogeneous biotitic matrix with kspar augen
54	664770	5183230	x			y	y	62	y		x		y	x	z					y	z	y				x		x	Pale grey leucocratic calc-silicate gneiss
90	657190	5168440	y			y	x	32	y		x		y	y						y	z		z						Mesocratic gneiss with hornblende seams.
92	659210	5167210	x				x	29	x					y	y									z	z				Fine grained flinty looking rock.
94	664250	5167920	x				x	28	y						y														Fine grained biotite quartzite, possibly mylonitic
96	664270	5167250	y				x	33	x				y											z					Biotite gneiss with granitoid lits
105	668460	5186130	x				x	55	y		x		y	x	y					y	y	y				y		x	Pale grey calc-silicate gneiss
Gneissic to lineated amphibolite (unit Hpba)																													
42-1	664160	5177070		y			x	36			x			y	z					y	z				z				Mesocratic amphibolite with epidote seams
45-2	665900	5179380	y				x	45	y		x			y						y	z						y		Very homogeneous, melanocratic amphibolite
49-1	661940	5178880	y	y			x	47			x			y							z								Amphibolite with 30% plagioclase augen
51-3	664590	5186190	y				x	41	y		x			y	z					y	z						y		Amphibolite gneiss with biotite seams
77-2	668590	5178850	y				x	39			x									z									Very homogeneous mesocratic amphibolite
79	666810	5179030	x	y			x	43			x			y						z									Homogeneous mesocratic amphibolite
97-1	664010	5167020	y	y			x	41	y		x			y							z						y		Strongly foliated amphibolite with quartz lenticles
Foliated augen granite gneiss																													
1	704550	5181970	x	x	y		x	20	x	z											z		z						Homogeneous granoblastic with biotite seams
22	658350	5176670	x	x			x	19							y						z		z			z			Brecciated pale grey granitoid gneiss
25	658480	5716310	x	x			x	22							x						z		z						Brecciated pinkish granitoid gneiss
32-2	661560	5177690	x	x			x	23	x														z			z			Slightly foliated granitoid with biotite fringing feldspar
46	665360	5180120	x	x			x	24	y	z													z			z			Faintly foliated grey granitoid
50-3	662190	5184590	x	x	y	y	x	19	y	z					y						z		z			z			Granoblastic granite gneiss
56-1	664870	5184760	x	x			x	17	y	z													z						Granoblastic gneiss poikilitic feldspar
107-1	670890	5185780	x	x			x	21	y	z											z								

Specimen	Easting (NTS 11K)	Northing	quartz	microcline	perthite	orthoclase	plagioclase	An(mol)	biotite	muscovite	hornblende	actinolite	garnet	epidote	chlorite	diaspore	sphene	apatite	carbonate	zircon	tourmaline	magnatite	ilmenite	pyrite	Description
Anorthosite (unit Hrra)																									
4-3	681150	5188270	z				x	53	y					y	y										Brecciated white foliated anorthosite
6	675580	5185940					x	56						y											Brecciated white anorthosite
58-1	676030	5186320					x	48						y											Brecciated, altered anorthosite
106	674820	5181450					x	51		z				y	z										Massive, slightly altered anorthosite
Mafic complex (unit Hm)																									
33-2	659770	5178540	y				x	56			x							z				z			Lineated, sericitized hornblendite
95-1	675210	5187120	y				x	62			x						z	z				z			Lineated, homogeneous mafic amphibolite
Cheticamp pluton (unit Hc)																									
21-1	658260	5176450	x				x	29	x					x	y				z	z		z			Granoblastic altered massive granodiorite
27-3	658460	5176290	x			y	x	32	x					y	y			z			z			z	Massive, biotite wrapped around feldspar crystals
34-2	659030	5177030	x			y	x	36	y		y			y				z	z			z			Faintly foliated, hornblende lineation
41-2	661940	5177870	x			y	x	31						y	x					z	z				Fractured, grey-green massive granodiorite
57-2	665140	5187020	x			x	x	33	y					y	z			z		z	z				Grey-green fractured granodiorite
65	659480	5179040	x			x	x	31	x					y	y				z			z			Fractured, altered pale pink granodiorite
70	659690	5175730	x			y	x	33	x					x	z			z		z					Fractured, grey-green massive granodiorite
91	658330	5167490	x			y	x	30						x	x			z	z		z	z			Grey-green, altered granodiorite with chlorite surfaces
Jumping Brook complex, volcanic member (unit Sjbv)																									
12-3	656070	5170990	x	x	y		4		z											z		z			Pale pink fine grained schist with 10 mm feldspar phenoclasts
27-2	658460	5176290	y				x	2							x	y	z					z			Finely laminated grey-green schist
32-1	661560	5177960	y				x	18		y	x			y			z				y				Lineated, greenish amphibolite schist
35-2	656050	5171710	x		x		y	4		y										z		z			White granular muscovite schist with phenoclasts
41-1	661940	5177870	y				x	19	y	z		x		y							z	z			Pale green actinolite schist
60	672780	5186420	x		x		y	8		y									z	z	z				Pink uniform schist with "knots" on surface
66	661920	5167230	y				x	26	x		x			x	z		z	z							Greenish black, strongly lineated amphibolite schist
69-1	665120	5184230	y				x	18	x			x		x			z				z	z			Pale grey-green, uniform actinolite schist
88	657240	5170210	x		x		y	0		y					y					z		z			Pale pink, fine grained schist with phenoclasts
99-2	661470	5166780	z				x	0				x	z	x	x							z			Fine grained, fractured, chocolate brown rock.

Appendix 1 (continued)

Specimen	Easting (NTS 11K)	Northing (NTS 11K)	quartz	microcline	perthite	orthoclase	plagioclase	An(mol)	biotite	muscovite	epidote	chlorite	garnet	staurolite	chloritoid	corclieniite	Al-silicate	apatite	carbonate	zircon	tourmaline	magnetite	ilmenite	pyrite	Description
Jumping Brook complex, pelitic gneiss and schist (unit Sjbp)																									
10	662830	5167000	x						y	x		y	y	x		x				z	z		z	z	Lustrous, silvery schists with 2 m "knots" on cleavage
19-2	657520	5174670	x							x		x	y		y					z			z		Grey green, fissile schist with rosettes of porphyroblasts
28-3	659720	5178290	x						x	x		y	x							z			z		Lustrous silver schist, 3 cm garnets
29-2	660010	5178780	x						x	x		y	y					z	z				z		Uniform pale grey, fissile schist
30-1	660830	5178800	x						x	x		y	y	x		y				z			z		Lustrous silvery schist with "knots" on surface
63-1	660250	5179390	x						y	x		y	y					z					z		Lustrous white schist with compositional layering
69-2	658470	5176390	x						x	x			y					z					z		Pale grey, fissile, muscovite-rich schist
98-1	663000	5167040	x						x	x		x	y	y						z				z	Pale grey, knotted schist
102	659410	5166270	x							x		x	y	y									z		Grey green, fissile, contorted mica schist
Jumping Brook complex, psammitic schist (unit Sjbg)																									
20-1	658290	5176410	x							x		x	y								y		z		Pale grey, gritty schist, micas not prominent
28-2	659720	5178530	x			y	y	25	x	x		x											z		Uniform, fine grained biotite-muscovite schist
29-3	660010	5178780	x			y			x	x		x									z		z		Quartz-rich schist with pebbles to 2 cm
31-2	661360	5178420	x			y			x		x	y									z		z	z	Fissile, fine grained, uniform quartz-mica schist
43-1	665270	5177940	x			y	y	20	x	x		y	y							z			z		Pale grey, faintly gneissic, quartz-mica schist
49-2	661940	5178080	x				y	20	x	y			y							z			z		Pale grey faintly gneissic, quartz mica schist
55-2	664720	5486680	x			y	y	22	x	y										z			z		Strongly fissile quartz-mica gneiss
Foliated muscovite biotite granite (unit SDg)																									
42-2	664160	5177070	x	x		x	x	19	x	y								z		z			z		Granular, white, foliated biotite muscovite granite
48	663160	5180020	x	x	z	x	x	21	y	y			y					z		z			z		Granular, white, foliated biotite muscovite granite
51-1	664590	5186190	x	x		x	x	19	y	y										z					Granular, pale-pink, biotite muscovite granite
53-3	664250	3184630	x	x	y		x	21	y	x			y							z			z		Weakly foliated muscovite biotite granite
95-2	664470	5167310	x	x		y	x	17	y	y										z					Granular, white, foliated biotite muscovite granite

Specimen	Easting (NTS 11K)	Northing (NTS 11K)	quartz	microcline	perthite	orthoclase	plagioclase	An(mol)	biotite	muscovite	epidote	chlorite	garnet	apatite	carbonate	zircon	tourmaline	magnetite	ilmenite	pyrite	fragments	matrix	Description
Lineated muscovite granite (unit Dl)																							
2	702950	5187030	x	x			x		z	y			z										Very coarse red muscovite granite with tiny garnet
24-2	659020	5178190	x	x			x	12	z	y		z	z										Graphic granite with aligned intergrowth
34-1	659030	5177030	x	x		z	x	4	z	y			z										Coarse red, lineated muscovite garnet granite
42-3	664160	5177070	x	x			x	10	z				z										Muscovite garnet aplite
62	665710	5187400	x	x		y	x	8		y		z											Coarse red muscovite garnet granite
Margaree pluton (unit Dm)																							
72	673670	5177210	x		x		x	29	z		z	y				z		z					Dyke of pink feldspar porphyry
73	673220	5184680	x		x		x	30				x		z		z		z					Megacrystic red biotite (chlorite) granite
74	668550	5167300	x		x		x	27	z			y				z		z					Megacrystic red granite, rapakivi feldspar
75	665100	5159060	x		x		x	27				y		z		z							Megacrystic red granite
76	669130	5178390	x		x		x	27				y				z		z					Megacrystic red granite
Salmon Pool pluton (unit Dsp)																							
11-1	656330	5170990	x		x		x	30	y			z		z		z		z					Massive, medium grained pink biotite granite dyke
43-2	665640	5178220	x	y	x		x	26	x	z			z	z		z		z					Faintly foliated pink biotite granite, muscovite on parting planes
93	660210	5167390	x	y	x		x	29	x	y				z		z		z					Massive, pink coarse biotite granite
100	667270	5166730	x	y	x		x	30	x			y		z		z		z					Massive medium grained pink biotite granite
109	671250	5187030	x		x		x	27	x			y		z		z		z					Sheared pink biotite granite
Fisset Brook Formation (unit DMFb)																							
11-2	656330	5170990					x	8		y	x	x						z					Fine grained vesicular purplish basalt
18-1	656040	5172170					x	11			x	x			z			z					Fine grained amydular black basalt
26-1	656910	5168620				y	x	9			x	x				z		z					Flow banded pink rhyolite
26-2	656910	5168620					x	6			x	x						z					Geenish amygdular basalt
35-1	656050	5171710	x			y	x	8	z		x	z				z							Pale pink rhyolite with contorted banding
38-1	656930	5169370					x	11			x	x			y		z	z					Brownish amygdular mafic volcanic
71	658470	5176390	x			y	x	4			x	y				z		z					Very deep red, faintly banded acid volcanic
80-2	657020	5167030					x	7			x	x						z					Purplish vesicular mafic volcanic
89	657310	5170590					x	11			y	x						z					Greenish amygdular volcanic
90-2	657190	5168440	x			y	x	8	z		y	y				z							Pale green aphanitic rhyolite
Cataclastic rocks																							
8	663200	5167100	x	x			x	26	y		y		y					z					Pale grey flinty rock with strong mylonitic banding
15-2	656270	5172210	x	y			y	?			z	y										x	Intensely banded, fine grained mylonitic rock
37	656850	5169090	x				x	22	y	y		z						z				x	Pink aphanitic rock with nebulous fragments
67-2	660970	5176160	x	x		z	x	26	y	y			y	y								x	Fine-banded greenish grey rock of mylonitic appearance

Appendix 2. Bulk-rock chemical composition of specimens from the Chéticamp region

Specimen	DF	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	H_2O	CO_2	P_2O_5	Ba	Cr	Cu	Ni	Pb	Rb	Sc	Sr	V	Yb	Zn	Zr	Total
Biotite gneiss and schist (unit Hpbp)																											
3-1	-0.84	63.6	1.12	14.3	0.9	6.1	0.13	3.49	2.11	4.2	2.88	1.1	0.0	0.21	670	45	9	30	7	50	19	300	200	4.3	110	90	100.1
45-1	2.74	52.4	1.55	17.0	3.7	5.7	0.17	4.13	6.68	3.7	2.52	1.2	0.0	0.34	450	21	7	20	6	40	23	290	230	4.2	80	110	99.3
21-2	-1.23	61.4	1.23	13.1	2.4	5.9	0.19	3.55	4.31	2.4	2.93	1.8	0.0	0.43	840	32	13	30	14	80	19	450	160	4.6	150	120	99.6
50-1	2.77	62.3	0.85	16.7	1.2	5.1	0.21	2.93	2.59	4.9	3.25	0.7	0.0	0.12	530	63	11	20	9	70	18	230	180	4.0	90	80	100.9
53-2	4.65	56.2	0.96	16.9	1.3	5.9	0.14	3.48	5.17	4.8	3.16	1.3	0.0	0.22	810	71	8	40	13	100	27	60	160	4.5	130	110	99.5
54	1.31	56.0	0.74	13.9	1.4	4.6	0.22	4.98	12.7	0.87	2.30	1.6	0.6	0.22	510	93	7	30	16	170	19	40	120	4.4	100	60	100.4
90	2.47	52.5	0.95	16.8	3.6	4.9	0.23	3.95	7.11	3.7	2.86	2.7	0.3	0.35	590	47	13	30	50	60	20	290	210	5.0	120	140	99.1
92	3.16	62.6	0.95	16.5	0.6	5.6	0.17	2.95	2.23	5.6	2.53	0.6	0.0	0.17	630	61	11	20	14	40	19	310	130	4.3	90	160	100.5
94	-4.68	87.9	0.12	6.43	0.1	1.0	0.06	0.58	0.85	2.3	0.58	0.3	0.0	0.01	170	47	2	10	1	30	11	110	130	2.1	30	150	100.2
96	+4.80	56.3	1.28	18.2	0.1	7.0	0.23	3.62	3.88	5.7	2.86	0.8	0.0	0.11	600	46	9	50	19	90	15	420	190	4.9	50	110	100.1
105	-2.56	65.5	0.36	10.6	1.9	6.1	0.20	3.13	8.39	0.91	0.99	1.1	0.7	0.25	570	62	15	10	11	120	30	70	220	3.8	100	70	100.1
Gneissic to lineated amphibolite (unit Hpba)																											
42-1		45.1	0.52	15.1	3.8	11.8	0.22	7.01	11.49	2.7	1.20	1.2	0.0	0.11	320	90	65	80	8	40	23	220	280	4.6	130	60	100.2
45-2		47.8	0.63	14.3	4.1	10.5	0.25	6.64	10.40	2.6	1.01	1.5	0.0	0.15	260	110	46	40	6	60	41	110	230	4.3	110	40	99.9
49.1		47.1	0.84	19.0	3.3	7.4	0.33	7.36	9.15	2.7	1.51	1.9	0.0	0.10	280	88	73	60	11	50	31	20	270	5.1	150	0	101.0
51-3		50.9	1.11	16.1	3.8	7.9	0.27	5.26	7.49	2.9	2.46	1.7	0.0	0.22	350	76	38	40	13	50	35	180	160	5.6	190	90	100.1
77-2		50.5	0.42	16.2	3.1	9.0	0.35	5.58	9.75	3.3	1.27	0.8	0.0	0.09	210	85	87	90	9	110	27	230	140	4.6	60	80	100.4
79		55.5	0.39	14.4	2.6	8.5	0.30	5.11	8.77	2.9	1.15	0.5	0.0	0.12	300	125	110	130	7	40	33	140	220	4.2	90	40	100.2
97-1		49.5	0.67	13.6	3.9	11.5	0.23	6.77	9.38	2.1	2.11	1.3	0.0	0.17	290	57	67	70	11	30	22	160	190	4.3	130	0	100.3
Foliated augen granite gneiss (unit Hpbg)																											
1		72.8	0.21	14.6	0.5	1.1	0.07	0.72	1.39	4.3	3.92	0.3	0.0	0.09	500	20	10	20	12	60	12	190	73	2.4	40	130	100.0
22		71.1	0.13	16.5	0.6	1.2	0.11	0.42	1.96	5.6	1.42	0.8	0.0	0.21	530	0	8	0	16	120	13	60	85	3.3	110	130	100.0
25		60.1	0.59	15.2	4.7	4.8	0.22	2.35	0.66	3.2	5.84	2.3	0.0	0.20	620	21	7	40	64	200	14	30	94	3.6	760	170	99.9
32-2		65.7	0.61	16.1	1.9	2.6	0.12	0.79	2.32	5.0	3.50	0.4	0.0	0.16	770	11	10	20	17	60	12	130	30	7.1	70	720	99.3
46		69.9	0.17	17.4	0.3	0.9	0.14	0.53	2.15	6.1	1.90	0.6	0.0	0.07	620	19	7	0	15	110	11	220	60	3.8	110	180	100.2
50-3		67.8	0.43	16.5	1.1	1.7	0.18	1.21	1.87	3.6	3.93	1.3	0.0	0.13	730	13	11	10	18	130	13	130	40	4.1	90	230	99.8
56-1		73.5	0.12	14.8	0.2	0.5	0.08	0.32	1.21	3.7	4.76	0.9	0.0	0.17	580	0	13	0	16	100	9	310	55	3.2	30	110	100.3
107-1		72.7	0.18	16.8	0.2	0.9	0.03	0.55	1.65	5.6	2.65	0.6	0.0	0.09	590	0	7	10	15	50	4	760	21	0	20	0	100.9
Anorthosite (unit Hrra)																											
4-3		54.4	0.14	26.3	0.9	0.4	0.03	0.80	9.29	6.1	0.73	0.8	0.4	0.06	160	0	7	20	3	0	0	110	18	0	0	0	100.6
6		53.1	0.15	26.3	1.6	0.7	0.02	0.65	9.01	5.6	1.21	1.1	0.3	0.07	320	0	3	0	1	10	6	230	10	2.1	50	0	99.6
58-1		54.9	0.09	26.8	1.2	0.2	0.05	0.56	7.73	5.7	0.92	0.9	0.1	0.07	230	0	11	10	0	0	4	160	24	0	0	0	100.0
106		55.1	0.15	26.9	1.1	0.7	0.02	0.71	7.13	5.5	1.43	0.8	0.2	0.04	450	0	0	0	0	0	0	420	14	0	0	0	99.8
Mafix complex (unit Hm)																											
33-2																											
95-1		45.9	0.96	18.5	0.6	7.7	0.16	11.4	10.6	2.4	0.34	1.9	0.0	0.14	10	370	14	200	1	10	22	280	140	2.2	40	0	100.9
Cheticamp pluton (unit Hc)																											
21-1		52.5	0.70	17.9	2.1	5.3	0.28	4.43	5.64	3.4	2.99	2.4	0.0	0.40	840	32	13	30	14	80	19	450	160	4.6	150	120	99.3
27-3		66.1	0.35	15.8	1.2	3.1	0.14	2.10	2.22	3.5	3.64	1.5	0.0	0.10	780	13	0	10	4	120	8	180	57	0	60	0	99.9
34-2		65.1	0.37	16.6	1.1	2.7	0.15	1.96	4.26	3.9	1.79	0.8	0.0	0.11	430	17	9	30	5	40	9	230	65	0	100	0	99.0
41-2		67.8	0.47	17.2	0.2	1.9	0.04	0.79	1.80	5.3	3.54	0.4	0.0	0.18	660	0	14	20	46	70	0	440	25	0	90	160	100.0
57-2		66.7	0.50	16.7	0.8	2.6	0.07	1.27	2.23	4.1	3.56	1.0	0.0	0.19	830	0	6	20	15	160	9	370	52	2.1	70	100	100.0
65		69.0	0.50	15.6	0.3	2.5	0.07	1.05	2.51	3.9	3.59	0.7	0.0	0.15	840	0	8	0	18	80	7	450	38	0	70	120	100.3
70		68.4	0.37	15.7	1.7	1.6	0.13	1.21	2.51	3.6	3.02	1.4	0.0	0.16	660	0	14	10	11	70	6	440	37	0	60	90	99.9
91		66.5	0.43	16.1	1.3	2.9	0.11	1.16	2.87	3.4	3.11	1.3	0.2	0.11	710	6	11	10	8	80	8	380	42	3.6	50	110	99.5
Jumping Brook complex, volcanic member (unit Sjbv)																											
12-3		72.0	0.35	15.2	1.3	0.3	0.03	0.41	0.22	5.0	4.50	0.6	0.0	0.08	560	0	10	20	7	100	6	60	24	4.4	0	320	100.1
27-2		51.8	1.75	16.6	4.0	7.0	0.21	5.72	6.44	2.7	0.59	3.7	0.0	0.30	50	58	29	30	4	0	28	320	250	4.6	100	60	101.0
32-1		48.5	1.04	17.0	1.0	7.6	0.19	8.50	10.4	2.6	1.34	1.9	0.0	0.11	270	250	100	100	11	40	30	410	190	3.7	60	0	100.5
35-2		72.1	0.28	15.1	0.9	1.6	0.07	0.53	0.19	5.2	4.33	0.4	0.0	0.11	430	0	9	10	5	130	7	80	18	5.3	50	280	100.8
41-1		49.5	1.53	15.8	1.7	7.1	0.22	7.63	9.76	2.9	1.13	2.3	0.0	0.25	110	160	23	80	8	20	36	310	280	3.2	120	0	99.8
60		73.5	0.21	14.8	0.9	0.3	0.06	0.65	0.23	4.9	4.13	0.6	0.0	0.09	520	0	6	0	9	110	4	60	22	4.1	0	300	100.4
66		49.1	1.47	16.1	1.3	6.8	0.19	7.76	10.03	3.4	0.98	2.6	0.0	0.22	370	50	80	30	11	30	8	440	190	4.3	90	60	99.9
69-1		53.4	0.93	16.7	1.0	6.0	0.20	7.43	7.09	4.9	1.22	1.5	0.0	0.21	220	150	3	180	12	40	34	220	120	2.6	70	40	100.8
88		72.6	0.31	15.3	0.6	0.6	0.10	0.73	0.35	5.1	3.78	0.4	0.0	0.11	490	0	7	0	0	130	8	90	20	5.3	0	380	100.4
99-2		46.8	0.84	15.5	3.6	7.0	0.20	8.53	12.4	2.7	1.34	2.9	0.0	0.10	10	330	36	260	3	0	20	660	200	4.1	70	0	100.0

Appendix 2 (continued)

Specimen	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	H_2O	CO_2	P_2O_5	Ba	Cr	Cu	Ni	Pb	Rb	Sc	Sr	V	Yb	Zn	Zr	Total
Jumping Brook complex, pelitic gneiss and schist (unit Sjbp)																										
10	52.8	0.99	22.1	1.1	7.6	0.34	3.35	1.46	2.3	3.92	3.6	0.0	0.16	420	110	100	60	14	160	23	80	130	4.6	110	80	99.8
19-2	55.4	1.17	21.1	1.0	8.6	0.18	2.27	0.66	0.87	3.71	4.3	0.0	0.36	810	97	19	40	23	160	20	140	180	5.3	160	100	99.7
28-3	62.3	0.93	18.2	1.1	4.7	0.33	2.25	1.43	2.1	3.81	2.5	0.0	0.15	620	57	12	30	37	110	150	150	110	4.0	1110	260	100.1
29-2	52.9	2.15	22.0	1.3	6.9	0.23	0.61	1.7	5.02	4.1	0.0	0.11		520	100	52	50	20	130	24	90	130	3.8	130	130	99.9
30-1	60.0	1.09	20.3	0.4	6.8	0.09	2.09	0.47	0.95	3.90	2.8	0.0	0.09	440	95	13	60	8	180	21	170	140	3.1	30	190	99.0
63-1	66.0	0.24	18.8	0.8	3.9	0.46	1.19	0.41	0.63	4.98	2.4	0.0	0.06	460	78	19	40	13	170	23	100	120	5.3	660	110	99.8
69-2	52.2	2.72	25.1	0.7	3.9	0.30	2.25	0.57	0.85	6.49	4.6	0.0	0.05	710	130	73	80	19	120	30	70	160	4.6	220	90	99.8
98-1	57.3	0.43	22.7	1.3	10.9	0.87	2.66	0.47	0.18	1.93	1.6	0.0	0.00	470	67	21	40	14	160	21	90	140	3.1	180	160	100.4
102	52.6	2.60	25.8	0.9	7.1	0.21	0.75	0.18	0.70	4.98	4.1	0.0	0.03	630	75	13	30	21	140	24	80	150	7.0	110	210	99.8
Jumping Brook complex, psammitic schist (unit Sjbg)																										
20-1	75.4	0.33	13.4	1.1	0.9	0.02	0.56	0.11	4.7	2.13	1.0	0.0	0.06	350	0	19	10	3	60	20	20	180	5.3	20	360	99.7
28-2	62.7	1.10	17.8	1.4	6.0	0.15	1.90	0.97	1.2	3.09	3.5	0.0	0.18	900	47	150	30	55	100	15	450	96	3.6	100	130	100.1
29.3	76.4	0.49	11.0	0.1	3.1	0.08	1.07	1.37	2.7	1.82	0.6	0.0	0.08	310	16	7	20	10	50	10	140	61	0	60	80	99.0
31-2	66.3	0.96	18.0	0.4	3.2	0.16	1.38	1.91	6.5	1.34	0.8	0.0	0.4	240	50	6	40	59	30	9	410	85	3.4	90	290	101.0
43-1	83.7	0.54	8.07	0.0	2.4	0.03	0.62	0.30	0.86	1.81	0.9	0.0	0.11	410	21	58	20	272	20	6	70	45	3.1	790	280	99.6
49-2	60.7	1.05	20.0	0.7	5.6	0.12	1.69	0.17	0.54	6.25	2.7	0.0	0.08	510	60	15	40	140	20	18	30	94	3.4	760	170	99.9
55-2	72.4	0.03	15.8	0.3	0.2	0.01	0.14	0.62	6.3	3.33	0.2	0.0	0.05	580	0	19	0	100	50	3	70	0	0	0	0	99.5
Foliated muscovite-biotite granite (unit SDg)																										
42-2	72.6	0.09	16.1	0.2	0.7	0.03	0.11	1.23	5.6	2.65	0.6	0.0	0.03	510	0	6	0	38	110	0	400	0	0	60	0	99.9
48	70.8	0.02	17.3	0.2	0.1	0.02	0.09	1.86	6.7	2.11	0.5	0.0	0.06	540	0	12	10	46	30	0	770	0	0	10	0	99.8
51-1	74.5	0.01	14.9	0.3	0.1	0.06	0.14	0.69	5.2	4.50	0.1	0.0	0.05	90	0	0	0	62	150	0	30	0	0	50	0	100.7
53-3	73.0	0.11	15.0	0.2	0.8	0.04	0.30	1.32	5.2	2.40	0.6	0.0	0.06	660	0	11	0	77	40	0	530	11	0	130	0	99.2
95-2	73.1	0.08	15.7	0.4	0.6	0.02	0.13	1.21	5.4	2.72	0.3	0.0	0.09	620	0	0	0	82	130	0	610	0	0	20	0	99.8
Lineated muscovite granite (unit Dl)																										
2	75.1	0.02	14.3	0.1	0.5	0.09	0.19	0.96	4.8	3.75	0.3	0.0	0.02	230	0	9	0	36	190	0	50	0	0	20	0	101.1
24-2	76.8	0.07	12.2	0.2	0.4	0.04	0.16	0.48	3.8	4.74	0.5	0.0	0.03	110	0	10	0	38	200	0	20	0	0	0	0	99.5
34-1	73.8	0.06	14.0	0.1	0.4	0.05	0.17	1.48	5.3	3.47	0.3	0.0	0.05	280	0	8	0	35	170	0	80	0	0	30	0	99.8
42-3	74.1	0.05	14.6	0.0	0.6	0.10	0.18	1.14	5.4	3.44	0.3	0.0	0.07	260	0	6	0	34	180	0	140	0	0	40	0	100.0
62	72.1	0.04	15.8	0.0	0.5	0.03	0.20	2.30	5.5	3.04	0.3	0.0	0.04	540	0	7	0	34	50	0	480	0	0	40	0	100.5
Margaree pluton (unit Dm)																										
72	70.8	0.50	13.7	1.1	1.7	0.06	0.54	0.53	3.2	6.48	0.6	0.0	0.16	710	0	7	20	6	130	6	60	34	3.4	0	240	99.5
73	72.1	0.63	13.1	1.3	1.2	0.05	0.61	0.49	3.1	6.45	0.7	0.0	0.17	610	0	10	20	8	150	7	70	31	3.3	0	230	99.9
74	73.6	0.47	13.0	1.2	1.9	0.07	0.53	0.55	3.3	5.02	0.5	0.0	0.16	630	0	19	10	11	100	9	50	29	5.1	10	290	100.3
75	71.7	0.55	14.1	1.6	0.9	0.04	0.69	0.43	3.4	5.43	0.8	0.0	0.16	540	0	7	10	6	120	3	60	2.9	0	0	120	99.8
76	74.0	0.61	12.7	1.9	0.6	0.4	0.76	0.44	3.3	5.29	0.9	0.0	0.15	500	0	5	10	7	130	6	10	37	3.7	0	290	100.8
Salmon Pool granite (unit Dsp)																										
11-1	74.3	0.22	13.1	0.7	0.6	0.03	0.36	0.41	3.5	5.91	0.6	0.0	0.07	430	0	8	10	10	220	0	60	8	4.1	10	210	99.8
43-2	75.7	0.20	12.6	0.9	0.7	0.02	0.15	0.22	3.2	5.03	0.6	0.0	0.04	640	0	5	10	17	90	5	60	0	7.0	10	260	99.6
93	75.1	0.20	13.1	0.6	0.6	0.02	0.31	0.14	3.1	6.13	0.5	0.0	0.09	270	0	10	20	11	220	3	20	11	3.3	10	80	100.0
100	73.1	0.24	13.6	0.8	0.6	0.04	0.39	0.83	3.6	5.56	0.3	0.0	0.06	480	0	0	20	0	240	0	70	0	0	0	110	99.3
109	73.7	0.24	13.6	0.9	0.5	0.03	0.22	0.61	3.7	6.11	0.4	0.0	0.09	410	0	0	10	6	190	2	50	6	2.4	20	190	100.1
Fisset Brook Formation (unit DMFb)																										
11-2	46.7	2.65	16.3	5.9	7.3	0.25	5.73	6.73	3.3	2.82	2.0	0.0	0.81	180	70	17	40	6	80	28	150	320	6.2	60	110	100.5
18-1	49.8	1.98	14.3	4.3	7.1	0.31	4.51	8.19	4.1	0.91	3.2	0.5	0.44	130	86	21	50	11	100	21	190	180	5.3	90	180	99.6
26-1 (a)	70.6	1.02	12.1	0.8	2.1	0.11	0.41	0.93	2.7	7.16	1.9	0.0	0.05	490	0	6	0	13	190	0	60	9	8.2	210	810	99.9
26-2	48.3	2.13	15.3	6.3	6.4	0.23	5.61	5.41	4.8	3.04	2.3	0.0	0.39	200	110	18	80	3	70	23	130	210	4.8	40	170	100.2
35-1 (a)	73.1	0.98	13.0	2.6	0.6	0.16	0.37	0.67	2.01	6.78	0.7	0.0	0.08	530	0	9	0	5	220	10	70	13	7.7	130	650	100.1
38-1	50-9	2.55	13.9	4.7	5.5	0.24	3.79	6.19	3.2	2.66	3.4	2.4	0.56	210	42	15	60	3	100	27	180	270	7.4	100	160	100.2
71	77.4	0.17	10.9	0.4	1.4	0.04	0.40	0.16	0.45	8.55	0.7	0.0	0.01	440	0	7	20	18	220	10	70	11	8.0	70	630	100.7
80-2	48.1	2.71	15.7	5.1	5.9	0.31	4.81	6.65	4.9	2.71	2.8	0.0	0.79	160	97	14	70	8	60	29	190	290	5.8	220	190	99.7
89	45.6	2.43	16.1	6.1	6.8	0.39	7.81	2.56	4.7	1.17	5.7	0.0	0.36	140	71	19	10	4	20	23	320	260	6.4	390	110	99.1
90-2	75.0	0.25	11.2	3.1	0.0	0.03	0.17	0.08	1.7	6.61	0.8	0.0	0.03	620	0	13	20	5	170	0	70	20	6.0	30	710	99.1
Cataclastic rocks (unit Mbr)																										
8	73.7	0.62	12.8	0.1	3.1	0.08	1.23	1.87	3.1	2.00	1.1	0.0	0.08	500	20	10	20	12	60	12	190	73	2.4	40	130	99.9
13-2	82.8	0.22	9.40	0.7	0.5	0.02	0.48	0.26	2.0	3.81	0.4	0.0	0.08	330	15	17	10	6	130	5	90	27	0	0	0	100.9
37	80.4	0.48	8.48	0.8	2.2	0.08	1.68	0.36	0.64	2.85	1.2	0.0	0.09	370	34	19	20	10	120	9	40	28	2.7	190	40	99.5
67-2	74.3	0.53	13.1	0.5	1.7	0.10	1.31	0.87	2.91	2.33	1.4	0.0	0.09	380	18	13	20	8	110	7	50	45	2.6	70	50	99.1

Appendix 3. Mineral chemical data for specimens from the Chéticamp region

Specimen	Mineral	SiO_2	TiO_2	Al_2O_3	Cr_2O_3	FeO(t)	MnO	MgO	CaO	Na_2O	K_2O	Total	n	Si	Ti	Al	Fe^3	Fe^2	Mn	Mg	Ca	Na	K	S	U
Biotite gneiss and schist (unit Hpbp)																									
45-1	biotite	36.49	2.36	14.90	0.02	19.52	0.22	11.58	0.17	0.24	9.55	95.05	6	2.80	0.14	1.35	0.00	1.25	0.01	1.33	0.01	0.04	0.94	22	8
	amphibole	41.51	0.71	11.70	0.04	20.22	0.31	9.25	11.70	1.75	1.30	98.49	10	6.33	0.08	2.10	0.00	2.58	0.04	2.10	1.91	0.52	0.25	46	16
	epidote	37.78	0.06	22.78	0.04	9.43	0.23	0.49	23.26	0.22	0.06	94.35	5	3.07	0.00	2.18	0.64	0.00	0.02	0.06	2.02	0.04	0.01	25	8
	plagioclase: ranges from An_{30} to An_{32} with slight reverse zoning																								
54	scapolite	46.34	0.01	27.30	0.04	0.05	0.01	0.17	16.46	3.21	0.31	93.90	4	7.14	0.00	4.96	0.00	0.01	0.00	0.04	2.72	0.96	0.06	50	16
	garnet (core)	37.55	0.08	19.86	0.05	23.87	4.75	3.10	9.90	0.00	0.01	99.17	7	2.99	0.01	1.87	0.13	1.46	0.32	0.37	0.85	0.00	0.01	24	8
	garnet (rim)	38.26	0.11	20.49	0.08	21.03	2.14	2.76	14.65	0.00	0.06	99.59	7	3.00	0.01	1.90	0.08	1.30	0.14	0.32	1.23	0.00	0.01	24	8
	biotite	36.59	1.98	15.79	0.02	17.73	0.22	11.78	0.07	0.04	9.77	93.93	16	2.81	0.12	1.43	0.00	1.14	0.01	1.35	0.01	0.01	0.96	22	8
	chlorite	26.54	0.14	17.43	0.05	24.29	0.42	15.34	0.04	0.00	0.09	84.34	8	2.91	0.01	2.25	0.00	2.22	0.04	2.50	0.01	0.00	0.01	28	10
	pyroxene	52.34	0.01	1.00	0.05	10.39	0.35	11.88	22.57	0.22	0.05	98.86	10	1.99	0.00	0.05	0.00	0.33	0.01	0.67	0.92	0.02	0.00	12	4
	epidote	38.85	0.09	28.34	0.05	5.71	0.12	23.11	0.15	0.15	0.06	96.72	11	3.03	0.01	2.60	0.36	0.01	0.01	0.03	1.93	0.02	0.01	25	8
	amphibole	42.61	0.69	12.57	0.06	16.14	0.32	9.50	12.21	1.15	1.57	96.91	14	6.46	0.08	2.25	0.00	2.05	0.04	2.15	1.98	0.34	0.32	46	16
	plagioclase: ranges from An_{57} to An_{88} in independent grains, and from An_{30} to An_{50} in intergrowths with microcline microcline: ranges from Or_{88} to Or_{93} with 21 counts Ba.																								
21-2	biotite	36.77	1.58	16.67	0.07	14.45	0.11	14.48	0.18	0.29	9.40	94.00	7	2.77	0.09	1.48	0.00	0.91	0.01	1.63	0.02	0.04	0.90	22	8
	amphibole	45.91	0.43	10.41	0.06	12.62	0.26	12.91	12.10	1.11	0.32	96.13	18	6.80	0.05	1.82	0.00	1.56	0.03	2.85	1.92	0.32	0.06	46	16
	chlorite 1	26.19	0.12	18.06	0.09	19.37	0.25	18.81	0.01	0.04	0.12	83.06	4	2.83	0.01	2.30	0.02	1.74	0.02	3.04	0.00	0.01	0.02	28	10
	chlorite 2	25.85	0.07	18.44	0.22	20.75	0.30	17.81	0.01	0.12	0.01	83.58	5	2.80	0.01	2.35	0.03	1.85	0.03	2.87	0.00	0.03	0.00	28	10
	microcline: ranges from Or_{89} to Or_{94} with 19 counts Ba.																								
	tourmaline	34.39	0.76	27.88	0.04	6.66	0.00	8.24	1.24	1.98	0.02	81.21	4	6.00	0.10	5.73	0.00	0.97	0.00	2.14	0.23	0.67	0.01	49	16
	plagioclase: ranges from An_{43} to An_{30} with reverse zoning.																								
Gneissic to lineated amphibolite (unit Hpba)																									
49-1	epidote	38.19	0.05	24.02	0.06	10.31	0.25	0.50	23.33	0.10	0.07	96.99	7	3.00	0.00	2.29	0.68	0.00	0.02	0.06	1.96	0.02	0.01	25	8
	amphibole	42.50	0.47	15.42	0.05	15.88	0.29	10.56	11.54	1.60	0.34	98.65	10	6.25	0.05	2.67	0.00	1.95	0.04	2.32	1.82	0.46	0.08	46	16
	plagioclase: ranges from An_{38} to An_{44} with reverse zoning.																								
Foliated augen granite gneiss (unit Hpbg)																									
107-1	muscovite	44.50	1.90	27.94	0.01	4.99	0.00	1.55	0.13	0.50	10.72	92.24	4	3.11	0.10	2.30	0.29	0.00	0.00	0.16	0.10	0.07	0.96	22	7
	biotite	37.28	2.83	15.74	0.01	21.05	0.44	9.60	0.08	0.09	9.64	96.71	4	2.82	0.16	1.41	0.00	1.33	0.03	1.09	0.01	0.01	0.93	22	8
	chlorite	27.44	0.07	21.27	0.05	30.91	0.14	6.45	0.04	0.01	0.55	86.93	4	2.99	0.01	2.73	0.00	2.81	0.01	1.05	0.01	0.00	0.08	28	10
	plagioclase: ranges from An_{16} to An_{23} in independent grains, An_{16} to An_{18} in intergrowths microcline: ranges from Or_{85-94} with a trace of Ba																								
56-1	muscovite	44.47	1.27	28.66	0.02	5.79	0.01	1.10	0.13	0.43	10.68	92.56	6	3.10	0.07	2.36	0.31	0.03	0.00	0.11	0.01	0.06	0.93	22	7
	epidote	38.71	0.20	24.22	0.05	11.14	0.33	0.33	23.41	0.18	0.05	98.62	2	3.01	0.01	2.22	0.73	0.00	0.02	0.04	1.95	0.03	0.01	25	8
	biotite	35.31	3.03	15.17	0.03	21.97	0.37	8.02	0.16	0.08	9.24	93.38	4	2.80	0.18	1.42	0.00	1.46	0.03	0.95	0.01	0.01	0.93	22	8
	microcline: ranges from Or_{84} to Or_{90} plagioclase: normally zoned from An_{31} to An_{19}, local inclusions of patch perthite with An_{2} to An_{5}																								
Cheticamp pluton (unit Hc)																									
27-3	epidote	37.97	0.05	23.50	0.04	11.58	0.60	0.18	23.44	0.16	0.05	97.57	9	3.00	0.00	2.19	0.77	0.00	0.04	0.02	1.98	0.03	0.01	25	8
	chlorite	25.81	0.12	19.26	0.05	20.14	0.87	18.94	0.00	0.00	0.01	85.20	3	2.73	0.01	2.40	0.11	1.67	0.08	2.99	0.00	0.00	0.00	28	10
	biotite	37.19	2.06	16.33	0.03	15.93	0.37	13.24	0.13	0.08	9.48	94.84	9	2.80	0.12	1.45	0.00	1.00	0.02	1.49	0.01	0.01	0.91	22	8
	orthoclase: ranges from Or_{58} to Or_{95} plagioclase: ranges from An_{40} to An_{31}, normally zoned with orthoclase rims																								
Jumping Brook complex, volcanic member (unit Sjbv)																									
13-2	epidote	37.96	0.03	23.64	0.03	12.75	0.13	0.82	21.49	0.28	0.09	97.22	6	2.99	0.00	2.20	0.84	0.00	0.01	0.10	1.82	0.04	0.01	25	8
	chlorite	25.96	0.09	20.27	0.05	22.41	0.40	17.20	0.02	0.13	0.01	86.54	4	2.73	0.01	2.52	0.03	1.95	0.04	2.70	0.00	0.03	0.00	28	10
	diaspore	4.5	-	80.9	-	0.6	-	-	0.9	-	-	86.9	3	0.04	0.00	0.93	0.00	0.05	0.00	0.00	0.01	0.00	0.00	3	1
	plagioclase: An_{0}																								
99-2	garnet	37.71	0.60	8.32	0.06	19.02	0.47	0.19	34.65	0.00	0.07	101.09	4	2.96	0.03	0.77	1.24	0.01	0.03	0.02	2.92	0.00	0.01	24	8
	epidote	38.49	0.02	23.00	0.06	11.24	0.08	0.20	23.72	0.03	0.05	96.89	7	3.05	0.00	2.15	0.75	0.00	0.01	0.02	2.01	0.01	0.01	25	8
	prehnite	42.89	0.04	20.49	0.05	3.59	0.01	0.16	27.17	0.14	0.04	94.58	10	3.02	0.00	1.70	0.21	0.00	0.00	0.02	2.05	0.02	0.00	22	7
	chlorite	27.82	0.05	19.01	0.09	19.70	0.41	19.71	0.11	0.00	0.01	86.90	3	2.88	0.00	2.32	0.00	1.67	0.04	3.04	0.01	0.00	0.00	28	10
	amphibole (core)	54.29	0.04	1.25	0.05	12.27	0.34	16.36	13.08	0.00	0.05	97.73	2	7.81	0.00	0.21	0.00	1.48	0.04	3.51	2.01	0.00	0.01	46	16
	amphibole (rim)	48.80	0.37	6.25	0.06	13.82	0.30	14.78	12.40	1.32	0.10	97.06	3	7.11	0.04	1.08	0.00	1.69	0.04	3.21	1.94	0.37	0.00	46	16
	plagioclase: Ab_{99-100}																								
Jumping Brook complex, pelitic gneiss and schist (unit Sjbp)																									
10	chlorite	24.46	0.20	19.68	0.05	26.45	0.98	13.77	0.00	0.00	0.07	85.66	5	2.68	0.02	2.54	0.08	2.34	0.09	2.25	0.00	0.00	0.01	28	10
	biotite	34.23	2.73	18.10	0.04	20.72	0.20	8.69	0.01	0.04	8.55	93.31	6	2.68	0.16	1.67	0.00	1.36	0.01	1.01	0.00	0.01	0.85	22	8
	garnet	37.45	0.09	20.39	0.05	30.99	5.42	2.24	3.88	0.01	0.00	100.52	8	3.01	0.01	1.93	0.04	2.04	0.37	0.27	0.33	0.00	0.00	24	8
	staurolite	27.51	0.63	53.87	0.03	14.36	0.62	1.17	0.00	0.26	0.00	98.45	12	3.90	0.07	9.00	0.00	1.71	0.07	0.25	0.00	0.07	0.00	47	-
	muscovite	46.87	0.34	34.68	0.00	1.75	0.04	1.21	0.14	0.70	9.73	95.46	6	3.11	0.02	2.71	0.00	0.10	0.00	0.12	0.01	0.09	0.82	22	7
	cordierite	46.76	0.05	32.96	0.04	11.98	1.06	5.47	0.01	0.52	0.02	98.87	6	4.91	0.00	4.08	0.00	1.05	0.09	0.86	0.00	0.11	0.00	36	-

n = number of analyses averaged. S = number of charges to which standard cell reduced
U = number of cations (adjusted by oxidation of Fe^{2+})

Appendix 3 (continued)

Specimen	Mineral	SiO_2	TiO_2	Al_2O_3	Cr_2O_3	FeO(t)	MnO	MgO	CaO	Na_2O	K_2O	Total	n	Si	Ti	Al	Fe^3	Fe^2	Mn	Mg	Ca	Na	K	S	U
19-2	chlorite	23.42	0.09	23.48	0.07	32.41	0.31	8.36	0.00	0.00	0.05	88.19	6	2.55	0.01	3.02	0.00	2.96	0.03	1.36	0.00	0.00	0.00	28	10
	garnet	36.76	0.03	21.64	0.02	29.78	8.64	0.88	2.31	0.02	0.00	100.08	7	2.98	0.00	2.07	0.00	2.02	0.59	0.11	0.20	0.00	0.00	24	8
	chloritoid	23.87	0.02	40.30	0.05	25.02	0.88	1.61	0.00	0.03	0.00	91.78	7	1.00	0.00	1.99	0.00	0.88	0.03	0.10	0.00	0.00	0.00	12	4
	muscovite	45.10	0.22	35.36	0.01	1.37	0.00	0.56	0.66	1.91	8.53	93.72	6	3.03	0.01	2.80	0.08	0.00	0.00	0.06	0.05	0.25	0.73	22	7
28-3	chlorite	23.40	0.13	21.64	0.05	25.19	0.03	13.38	0.00	0.07	0.04	83.93	12	2.59	0.01	2.82	0.00	2.25	0.00	2.29	0.00	0.01	0.01	28	10
	biotite	34.75	1.58	18.02	0.03	19.60	0.02	9.34	0.13	0.27	8.54	92.28	10	2.73	0.09	1.67	0.00	1.29	0.00	10.09	0.01	0.04	0.86	22	8
	muscovite	45.36	0.39	35.16	0.02	1.16	0.01	0.94	0.12	1.17	9.52	93.85	13	3.05	0.02	2.79	0.04	0.03	0.00	0.09	0.01	0.15	0.82	22	7
	garnet (core)	36.42	0.29	20.02	0.10	27.48	6.74	1.47	5.88	0.26	0.02	98.68	6	2.97	0.02	1.92	0.13	1.75	0.47	0.18	0.51	0.04	0.00	24	8
	garnet (rim)	37.42	0.11	20.52	0.03	34.74	0.46	2.63	4.19	0.04	0.00	100.14	6	3.00	0.01	1.94	0.04	2.29	0.03	0.32	0.36	0.01	0.00	24	8
30-1	biotite	36.45	1.47	19.68	0.04	19.53	0.00	10.21	0.05	0.21	8.00	95.64	9	2.73	0.08	1.74	0.00	1.22	0.00	1.14	0.00	0.03	0.76	22	8
	chlorite	24.08	0.12	22.99	0.04	25.16	0.02	14.29	0.00	0.00	0.04	86.74	9	2.56	0.01	2.89	0.00	2.24	0.00	2.27	0.00	0.00	0.01	28	10
	muscovite	45.73	0.31	36.27	0.01	0.88	0.00	0.69	0.10	1.95	8.22	94.16	9	3.05	0.02	2.85	0.00	0.05	0.00	0.07	0.01	0.25	0.70	22	7
	staurolite	27.62	0.45	54.18	0.03	14.36	0.01	1.60	0.00	0.22	0.00	98.47	8	3.90	0.05	9.02	0.00	1.70	0.00	0.34	0.00	0.06	0.00	47	-
	garnet	37.31	0.10	20.30	0.07	35.94	0.92	2.41	3.54	0.00	0.00	100.59	6	3.00	0.01	1.92	0.07	2.35	0.06	0.29	0.31	0.00	0.00	24	8
Jumping Brook complex, psammitic schist (unit Sjbg)																									
20-1	muscovite	45.76	0.22	36.03	0.01	1.26	0.00	0.66	0.14	1.65	8.80	93.27	9	3.05	0.01	2.83	0.01	0.06	0.00	0.07	0.01	0.21	0.75	22	7
	chlorite	23.21	0.13	20.87	0.07	31.60	0.16	9.08	0.00	0.00	0.00	85.12	5	2.63	0.01	2.78	0.00	2.99	0.02	1.53	0.00	0.00	0.00	28	10
	tourmaline	34.56	0.58	30.27	0.05	8.80	0.02	5.32	0.35	1.94	0.05	81.94	3	6.00	0.08	6.19	0.00	1.28	0.00	1.38	0.07	0.65	0.01	49	16
	garnet (core)	36.29	0.16	20.10	0.06	26.72	10.92	0.76	3.43	0.00	0.00	98.44	6	3.01	0.01	1.96	0.00	1.85	0.77	0.09	0.30	0.00	0.00	24	8
	garnet (rim)	36.34	0.09	20.07	0.05	34.70	3.98	1.22	2.18	0.00	0.05	98.68	6	3.01	0.01	1.96	0.02	2.38	0.28	0.15	0.19	0.00	0.01	24	8
Jumping Brook complex (psammitic schists, continued)																									
28-2	biotite	35.56	2.77	16.71	0.03	22.24	0.27	7.59	0.09	2.05	9.55	94.86	6	2.77	0.16	1.53	0.00	1.45	0.02	0.88	0.01	0.01	0.95	22	8
	chlorite	24.04	0.16	19.92	0.05	28.64	0.56	11.64	0.01	0.00	0.09	85.11	5	2.68	0.01	2.62	0.00	2.67	0.05	1.94	0.00	0.00	0.01	28	10
	muscovite	46.61	1.01	32.98	0.00	2.26	0.01	1.41	0.12	0.59	10.82	95.81	6	3.10	0.05	2.58	0.11	0.01	0.00	0.14	0.01	0.08	0.92	22	7
	potash feldspar: ranges from Or_{88} to Or_{96} plagioclase: ranges from An_{24} to An_{29}; some grains An_4																								
43-1	garnet	37.40	0.06	20.42	0.04	37.01	1.80	2.60	1.36	0.00	0.00	100.69	6	3.01	0.00	1.94	0.03	2.46	0.12	0.31	0.12	0.00	0.00	24	8
	biotite	34.71	2.54	17.85	0.04	22.20	0.02	6.76	0.10	0.05	9.05	93.32	6	2.74	0.15	1.66	0.00	1.47	0.00	0.79	0.01	0.01	0.91	22	8
	chlorite	25.46	0.24	18.37	0.04	30.70	0.04	9.30	0.01	0.00	0.13	84.29	5	2.89	0.02	2.45	0.00	2.91	0.00	1.57	0.00	0.00	0.02	28	10
	muscovite	46.00	1.00	35.42	0.01	1.17	0.00	0.79	0.13	1.05	9.74	95.31	8	3.05	0.05	2.77	0.00	0.07	0.00	0.08	0.01	0.14	0.83	22	8
	potash feldspar: ranges from Or_{84} to Or_{97} plagioclase: ranges from An_{16} to An_{28}, perthitic patches in microcline An_2																								
Foliated muscovite-biotite granite (unit SDg)																									
42-2	muscovite	45.42	1.53	32.02	0.01	1.83	0.00	1.32	0.21	0.36	10.53	93.43	6	3.10	0.08	2.58	0.06	0.05	0.00	0.13	0.15	0.07	0.92	22	7
	biotite	36.02	3.40	16.97	0.01	22.21	0.26	7.63	0.14	0.13	9.73	96.50	6	2.76	0.20	1.53	0.00	1.42	0.02	0.87	0.01	0.02	0.95	22	8
	microcline: ranges from Or_{84} to Or_{92} plagioclase: ranges from An_{16} to An_{28}, perthitic patches in microcline An_2.																								
Lineated muscovite granite (unite Dl)																									
24-2	biotite	34.96	3.12	16.07	0.02	27.27	0.59	4.50	0.01	0.25	8.76	95.55	3	2.77	0.19	1.50	0.00	1.81	0.04	0.53	0.00	0.04	0.89	22	8
	chlorite	24.14	0.14	19.09	0.02	35.78	0.91	6.43	0.00	0.05	0.02	86.58	5	2.76	0.01	2.57	0.00	3.42	0.09	1.09	0.00	0.01	0.00	28	10
	muscovite	44.90	0.66	29.24	0.01	6.26	0.03	1.24	0.18	0.44	10.71	93.67	8	3.09	0.03	2.37	0.36	0.00	0.00	0.13	0.01	0.06	0.94	22	7
	garnet	36.68	0.03	19.51	0.01	19.74	15.70	0.49	6.48	0.02	0.00	99.16	8	3.00	0.00	1.88	0.11	1.24	1.09	0.06	0.61	0.00	0.00	24	8
	potash feldspar: both microcline and orthoclase Or_{93} to Ab_7 plagioclase: $An_{12}Ab_{87}Or_1$, minor islands of Ab_{92} to Ab_{99} in potash feldspar																								
Margaree pluton (unit Dm)																									
72	chlorite	24.74	0.09	17.21	0.05	32.45	0.78	11.17	0.00	0.18	0.00	86.67	6	2.75	0.01	2.26	0.26	2.76	0.07	1.85	0.00	0.04	0.00	28	10
	potash feldspar: host of perthite, and independent grains: Or_{98} plagioclase: flames in perthite An_{29}, patches in perthite An_1 to An_3																								
76	chlorite	26.27	0.09	16.92	0.06	26.47	0.32	14.70	0.03	0.03	0.02	85.51	3	2.88	0.01	2.18	0.05	2.38	0.09	2.40	0.00	0.01	0.00	28	10
	potash feldspar: host of perthite and independent grains Or_{98} plagioclase: flames in perthite An_{26}, patches in perthite An_1																								
Salmon Pool granite (unit Dsp)																									
100	chlorite	26.35	0.26	15.17	0.06	32.82	0.47	11.07	0.02	0.00	0.12	86.34	3	2.96	0.02	2.01	0.04	3.05	0.05	1.85	0.00	0.00	0.02	28	10
	biotite	36.85	4.12	11.90	0.01	24.79	0.43	8.69	0.01	0.02	9.07	95.89	6	2.88	0.24	1.10	0.00	1.62	0.03	1.01	0.00	0.00	0.91	22	8
	potash feldspar: host of perthite Or_{90}; cores of independent grains Or_{84}, rims Or_{88} plagioclase: independent grains An_{18} to An_{35}, flames in perthite An_{10} to An_{17}																								
Cataclastic rocks (unit Mbr)																									
8	biotite	34.46	3.12	17.84	0.01	20.13	0.23	7.47	0.34	0.25	9.12	92.97	6	2.71	0.19	1.66	0.00	1.33	0.02	0.88	0.03	0.04	0.92	22	8
	chlorite	26.37	0.34	17.36	0.05	28.55	0.33	11.75	0.01	0.02	0.33	85.11	6	2.93	0.03	2.28	0.00	2.66	0.03	1.95	0.00	0.00	0.05	28	10
	muscovite	46.70	1.03	33.26	0.01	1.60	0.00	1.35	0.12	0.56	10.64	95.27	6	3.12	0.05	2.62	0.02	0.07	0.00	0.13	0.01	0.07	0.91	22	7
	potash feldspar: ranges from Or_{93} to Or_{97} plagioclase: ranges from An_{21} to An_{24}																								

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